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Active tectonics of the Alpine–Himalayan belt: the Aegean Sea and surrounding regions

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Summary. New fault plane solutions, *Landsat* photographs, and seismic refraction records show that rapid extension is now taking place in the northern and eastern parts of the Aegean sea region. The southern part of the Aegean has also been deformed by normal faulting but is now relatively inactive. In northwestern Greece and Albania there is a band of thrusting near the western coasts adjacent to a band of normal faulting further east. The pre-Miocene geology of the islands in the Aegean closely resembles that of Greece and Turkey, yet seismic refraction shows that the crust is now only about 30 km thick beneath the southern part of the sea, compared with nearly 50 km beneath Greece and western Turkey. These observations suggest that the Aegean has been stretched by a factor of two since the Miocene. This stretching can account for the high heat flow. The sinking slab produced by subduction along the Hellenic Arc may maintain the motions, though the geometry and widespread nature of the normal faulting is not easily explained. The motions in northwestern Greece and Albania cannot be driven in the same way because no slab exists in the area. They may be maintained by blobs of cold mantle detaching from the lower half of the lithosphere, produced by a thermal instability when the lithosphere is thickened by thrusting. Hence generation and destruction of the lower part of the lithosphere may occur beneath deforming continental crust without the production of any oceanic crust.

1 Introduction

Most Earth scientists now agree that plate tectonics provides a satisfactory description of oceanic deformation. However, the deformation of continental regions is obviously more complicated than that of the oceans, and the general rules governing continental deformation are not yet agreed. At present there are two extreme points of view. One is that the deformation can usefully be described by the motion of a number of small plates moving in a rather complicated way (McKenzie 1970, 1972; Nowroozi 1972). The other is a modification of the earlier ideas of Argand (1924) put forward by Molnar & Tapponnier (1975), who argued that continental regions behave plastically and important deformation is distributed over wide regions rather than concentrated on a few large faults. Neither description is yet sufficiently precise to allow reconstructions to be carried out with any confidence, and hence the relationship of the observed motions of the relevant major plates to the geological history of the fold belts is unclear. The purpose of this and later papers is to examine interesting parts of the Alpine–Himalayan zone in some detail to attempt to understand the processes now taking place. This zone is the only continental region where large-scale shortening is now taking place. The major sources of information are the distribution and mechanisms of earthquakes, *Landsat* photographs of the land areas and oceanographic observations of submarine ones. Detailed maps of surface breaks of larger earthquakes are extremely important, but unfortunately few such reports exist. The principal interest in studies of this kind is as a framework for understanding past continental deformation and its evolution with time. The plate model proposed by McKenzie (1972) (hereinafter I) for the deformation of the Aegean and surrounding areas is shown in Fig. 1.

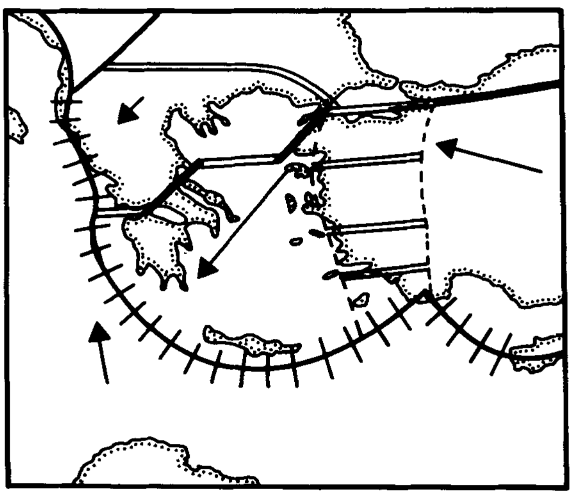

Figure 1. Sketch of the plate boundaries and motions in the Aegean area from McKenzie (1972). Extensional plate boundaries are shown by a double line, transform faults by a single heavy line and boundaries where shortening is occurring, by a solid line crossed by short lines at right angles. The fine dashed lines in western Turkey mark the E and W boundaries of the region of active E–W normal faults. The arrows show the direction of motion relative to Eurasia and their lengths are proportional to the magnitude of the velocity.

This model explained the majority of earthquake mechanisms in the region but has been criticized by a number of authors (Mercier *et al.* 1976; Bingöl 1977; Crampin & Ucer 1975). The principal modification required by extensive investigations of Mercier *et al.* (1976) is in central Greece, where they found no evidence of a transform fault. A different model has been suggested by Makris (1976), who believes that the deformation of the region is the surface expression of a mantle plume.

The locations of the geographical names used in this paper are shown in Fig. 18.

2 Data reduction

The hypocentres used in this study are those calculated by the United States Coast and Geodetic Survey and by the National Ocean and Atmospheric Administration. The accuracy of these locations increases with both the magnitude and the date, since it depends primarily on the number of stations reporting *P*-wave arrival times. The epicentres of large shocks ($m_b \geqslant 5.5$) are probably accurate to about 10 km, though there is some evidence of a systematic displacement to the north by about the same amount. Small shocks ($m_b < 5.0$) are much less accurately located. The depth control is poorer for all magnitudes because *pP* is rarely recorded and the control from close stations is inadequate. For small shocks especially the depth error can be as large as 75 km. Recent locations from a local network in Greece have shown that several shocks located by NOAA at depths below 70 km are in fact much shallower (Berckhemer, private communication). For these reasons complicated models of the shape of the subducted slab beneath the Hellenic Arc (Papazachos 1973, 1976a, b) should be viewed with caution. The same remarks apply to the distribution of small shocks along the Mediterranean Ridge.

The focal mechanisms used here are those from I together with a number of new solutions determined from long-period vertical WWSSN records read by the author. Dewey (1976) has published a solution for one of the shocks, 1970. 3.28 2102 GMT, which agrees excellently with that in Fig. 8(a). Ritsema (1974) has also published 17 solutions for shocks whose mechanisms are included in this paper. He does not show the polarity observations, and read only a few of the seismograms himself. He divides the solutions into four groups on the basis of their quality. The agreement between Ritsema's solutions and those below is excellent for about half the shocks, but there is no correlation between his estimate of the quality of the solutions and the extent of the agreement. Where his solutions disagree with those below his nodal plane often passes on the wrong side of some readings whose waveform suggested they were close to a node and whose polarity must be read with care. It seems likely that Ritsema used the wrong polarities for some of these stations.

Solutions for a number of the shocks in the Aegean have also been determined by Alptekin (1973) (see Bingöl 1976). His solutions have a much greater component of strike slip motion than those below, probably because he uses a lower velocity at the hypocentre. This has the effect of moving all the stations toward the centre of the projection (see I), which often increases the strike slip component. It is not clear what source velocity should be used for a crustal shock. If the polarity observations use wavelengths much greater than the crustal thickness (20 s or greater period) then a mantle velocity should be used. If, however, the wavelength is much less than the crustal thickness (periods 1 s or less) then the true source velocity is the relevant one. The problem with the long-period WWSSN records is that the instrumental response is greatest at about 15 s and the onset of the motion contains higher frequencies. Hence the periods involved are intermediate, and it is not clear what velocity to use. A study of this problem using synthetic seismograms would be of considerable interest. The procedure used here and in I was to use a mantle velocity of 7.8 km/s for all solutions. If two orthogonal planes could not be fitted to the observations the source velocity was reduced to 6.8 km/s and a new attempt made, which was in all cases successful.

The best satellite photographs for tectonic purposes are those taken with a low Sun angle in the late autumn and winter (October to December). If, however, snow is present, the automatic contrast control causes the snow-free areas to be underexposed (the opposite problem occurs when water covers an appreciable part of the scene). The elevation was used as a guide to the likelihood of snow cover when selecting the photographs, which were combined into mosaics. Before their edges were cut off latitude–longitude crossings were marked with a small cross. To make overlays showing the seismicity and fault plane solutions the completed mosaic was placed on an $X-Y$ digitizing table, the positions of the latitude–longitude crossings punched on to paper tape, and conic projection fitted to the points by least squares. The difference between the positions of the latitude–longitude crossings calculated from this projection and those on the mosaics is given in the figure captions, and is in most cases less than 5 km. The worst error was 7.3 km for the mosaic covering Greece, Albania and Yugoslavia (Fig. 17), and was caused by the absence of photographs in western Bulgaria and eastern Yugoslavia.

The same program can be used to make accurate overlays for the $1{:}10^6$ air approach maps, with errors of about 0.2 km. Unfortunately the air approach maps do not use the same projection as the satellite photographs, and therefore the overlays produced for these maps do not fit the mosaics as well as those produced by the procedure described above.

3 The Hellenic Trench

The Hellenic Trench consists of a number of deep basins between Kefallinia and Rhodes (Fig. 2), and is considerably shallower than the trenches of the western Pacific. There is an obvious difference between the basins west of Crete, which are little elongated along the arc, and the narrow linear structures called the Pliny and Strabo trenches, east of Crete. The seismicity is mostly seaward of the islands, and all of the larger shocks in Fig. 3 lie between the deepest part of the trench and the islands. A few small shocks may lie as far south as the Mediterranean Ridge, the swell south of the deep basins in Fig. 2, but because the events are small their locations are poor. At the northwestern end of the arc, where the sea-floor shoals, the seismically active zone runs into northwestern Greece (see also Fig. 14). Both the small-scale bathymetry (Stride, Belderson & Kenyon 1977) and geological mapping on land (British Petroleum 1971) suggest that the deformation is taken up by a number of right-handed strike slip faults which connect the deformation of northwestern Greece to the Hellenic Arc. It is surprising that none of the five fault plane solutions in this area in Fig. 3 show any evidence of strike slip movement. At the eastern end the activity is continuous with the normal faulting zone in western Turkey (Fig. 6). In the central region, however, there is little activity north of the Arc, and the few shocks which have occurred have all been small. Hence the southern part of the Aegean Sea is moving as a relatively rigid block compared with the surrounding zones. However, the low level of seismic activity, the seismic reflection records of Jongsma *et al.* (1977), and the existence of a number of active faults in the Peloponnese (Dufaure 1965) all show that slow deformation must none the less be occurring.

Table 1. The magnitude is the body-wave magnitude given by NOAA. Mechanisms of events before 1970 are contained in McKenzie (1972), listed in Table 3. If the depth is followed by an m the fault plane solution is that for a shock in the mantle, if followed by a c the solution is that for a shock in the crust with a velocity of 6.8 km/s.

Year/Month/Day	Time h m s	Lat °N	Long °E	Depth (km)	Magnitude	Figure
1975.9.22	00 44 57.7	35.29	26.23	63m	5.3	6 (b)
1975.4.30	4 28 56.9	36.18	30.77	56m	5.6	4 (f)
1975.4.4	5 16 16.2	38.09	21.98	53m	5.4	16 (d)
1975.3.27	5 15 06.2	40.42	26.14	5c	5.7	9 (k)
1973.11.29	10 57 42.7	35.18	23.80	26m	5.7	4 (e)
1973.11.4	15 52 11.7	38.90	20.44	8m	5.8	16 (c)
1973.1.5	5 49 17.5	35.81	21.84	33m	5.3	4 (d)
1972.9.17	14 07 15.6	38.28	20.34	33m	5.6	4 (c)
1972.9.13	4 13 20.5	37.93	22.39	83m	6.0	6 (a)
1972.5.4	21 40 00.9	35.12	23.61	46m	5.9	4 (b)
1972.3.14	14 05 45.8	39.28	29.42	33m	5.4	9 (j)
1971.5.25	5 43 27.0	39.03	29.74	24c	5.8	9 (i)
1971.5.12c	12 57 24.8	37.58	29.60	33c	5.4	9 (h)
1971.5.12b	10 10 37.2	37.53	29.72	33c	5.5	9 (g)
1971.5.12a	6 25 13.0	37.59	29.76	23c	5.5	9 (f)
1971.1.3	23 18 41.3	34.67	26.34	32m	5.2	4 (a)
1970.8.19	2 01 53.1	41.10	19.77	33c	5.2	16 (b)
1970.4.23	9 01 24.7	39.13	28.70	18m	5.2	9 (e)
1970.4.19	13 29 36.4	39.06	29.83	20c	5.4	9 (d)
1970.4.16	10 42 18.8	39.03	30.00	9c	5.5	9 (c)
1970.4.8	13 50 27.2	38.43	22.66	17m	5.8	16 (a)
1970.3.28b	23 11 44.0	39.22	29.52	37c	5.2	9 (b)
1970.3.28a	21 02 23.4	39.20	29.57	20c	6.0	9 (a)

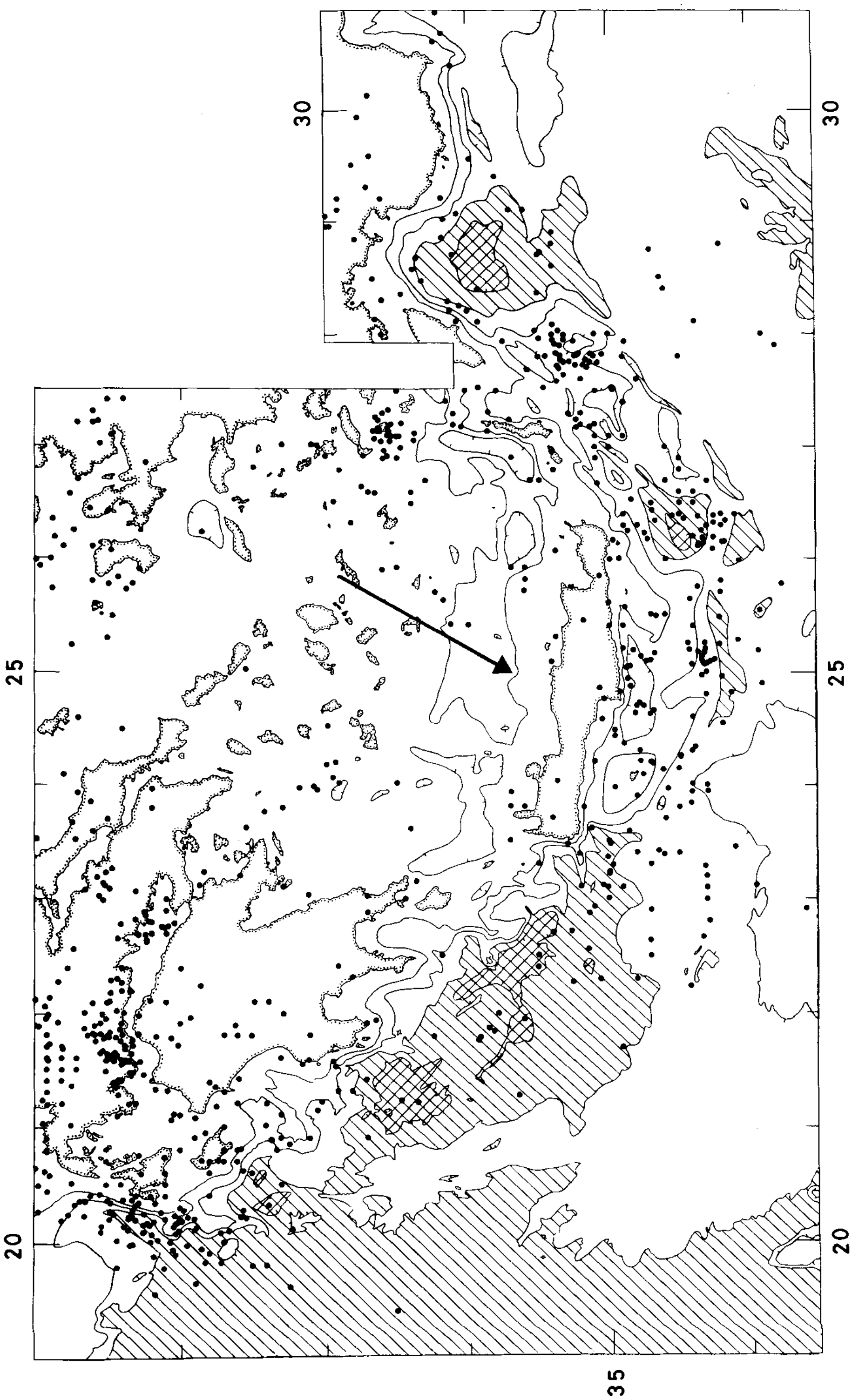

Figure 2. Earthquake epicentres for shocks with depths of 50 km or less between 1961.1.1 and 1975.12.31 given by the USCGS and NOAA. Bathymetric contours at 1-km intervals from Morelli, Gantar & Pisani (1975), Morelli, Pisani & Gantar (1975) and Wright *et al.* (1975). Depths greater than 3 km ruled, greater than 4 km cross-hatched. The heavy arrow shows the direction of relative motion (211° E) between the southern Aegean region and Africa determined from the fault plane solutions in Fig. 3.

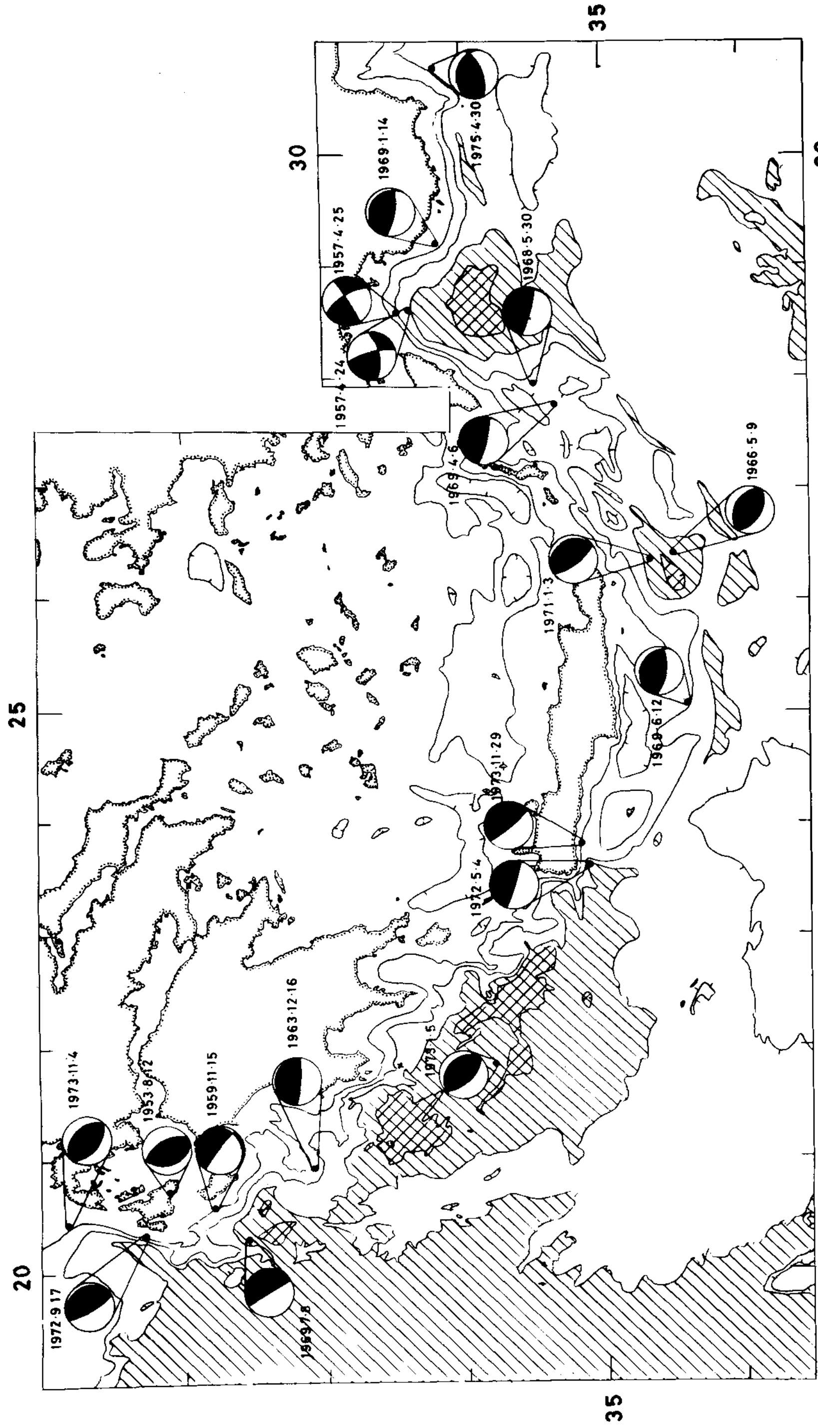

Figure 3. Lower-hemisphere projections of fault plane solutions from shocks associated with the Hellenic Arc (see Fig. 12 for those in the Aegean). The quadrants with compressional first motions are shaded. Contours at 1-km intervals (see Fig. 2).

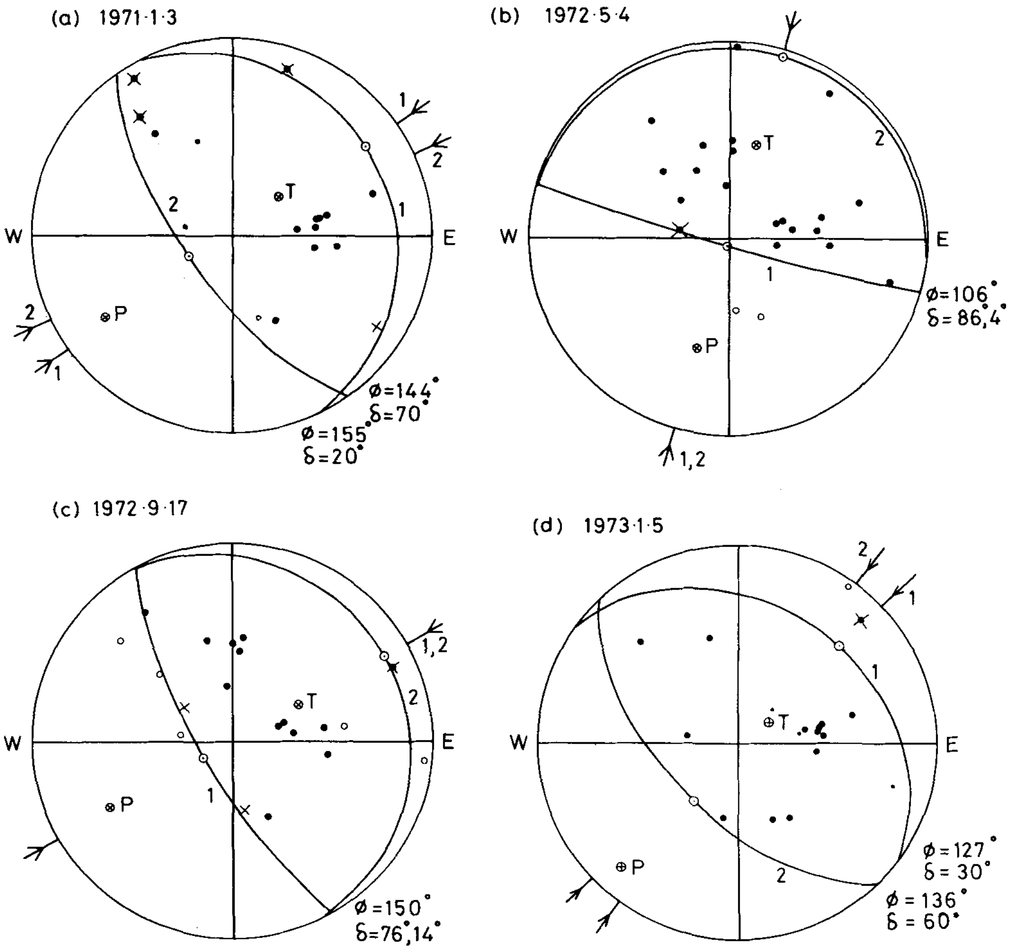

Figure 4. Polarity observations for the fault plane solutions shown in Fig. 3 which are not given in I. All mechanisms in this and other figures are equal-area projections of the lower hemisphere of the focal sphere. Long-period polarity observations obtained from the vertical long-period record at each station are shown as dots, open if the polarity was dilatational (first motion downwards), solid if it was compressional (first motion upwards). If the polarity was not completely clear, or if it had to be determined from short-period records it is shown as a small dot or circle. It is sometimes possible to recognize that a station is close to one of the nodal lines of the solution from the character of the long-period waveform itself, and such observations are shown as a dot or a circle if the polarity could be read, or as a simple cross if this was not possible. The focal planes are labelled 1 and 2 arbitrarily, and which is the fault plane cannot be decided from the polarity observations. If plane 1 (2) is the fault plane then the slip vector is the line labelled 1 (2). All solutions are placed in the mantle.

The mechanisms in Fig. 3 are reasonably consistent. All have a steeply dipping plane striking NW–SE. All shocks whose solutions were obtained by reading the WWSSN seismograms (1963.12.16 onwards) have a shallow plane dipping beneath the arc and a steeper one dipping SW. If these steep planes are the auxiliary planes, the mean slip vector has an azimuth of 211°, shown as an arrow in Fig. 2. This direction marks the relative motion between the southern Aegean and Africa and is approximately parallel to the narrow slots in the sea-floor known as the Pliny and Strabo trenches east of Crete. Hence the eastern part of the Hellenic Trench must be dominantly a transform fault. It is surprising that the

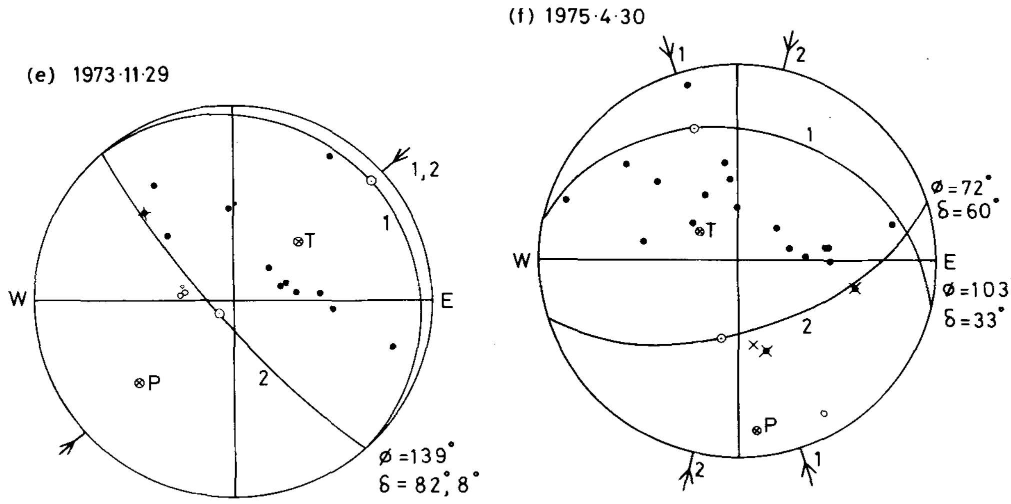

Figure 4. (e–f).

fault plane solutions in this area are almost all thrusts, though none of the epicentres are on the supposed transform faults. Only 1957.4.24 and 1957.4.25 show strike slip motion, and neither solution is well constrained (see I). Presumably the transform faults join a number of short segments of trench.

A surprising feature of Fig. 2 is the low level of seismic activity between Crete and the Peloponnese. This part of the arc is normal to the direction of relative motion and should be the most active. Perhaps this gap will be filled by a series of large shocks or perhaps the seismic activity of arcs depends on the thickness of the sediments on top of the plate being subducted. Certainly the frequency of large shocks along the Makran Coast in the northwest Indian Ocean, where thick sediment is also being subducted, is less than that of the Pacific Arcs.

The relationship between the surface deformation and the locations of intermediate-focus earthquakes is confused by errors in the published depths. The depths used in plotting Fig. 5 are those given by the USCGS and NOAA. Similar plots have been published by Papazachos (1973, 1976a) which include shocks before 1961. Unfortunately the depths of all these events are of uncertain reliability (see Section 2). Some of the small aftershocks of large, shallow, normal faulting events, for instance 1975.3.27, are claimed to be below a depth of 50 km, which seems unlikely. The seismic belts crossing continental regions in the Mediterranean and Middle East all contain a few small shocks located by the USCGS and NOAA at depths between 50 and 100 km, but wherever detailed microearthquake studies have been carried out these depths have not been confirmed. Until good depth determinations are available for the shocks in Fig. 5 the geometry of the intermediate-depth earthquake zones will remain uncertain.

The depths of shocks below 100 km is frequently known more accurately than that of shallower shocks. The majority are related to the subduction now occurring along the Hellenic Arc and increase in depth with distance from the thrust zone. If they lie within a slab produced by subduction at the Arc in the direction marked by the arrow in Fig. 2, the dip of the slab must be greater beneath Greece than beneath the southern part of the Aegean. There is, however, no reason to believe that the direction of subduction has remained unchanged for long periods. The extension which produced the deep basins north and east of

(a)

Plate 1. *Landsat* images of regions 2 (a), 821 300 814 25N and 811 150 820 45G and 3 (b), 811 150 820 45G and 811 150 821 05G, shown in Fig. 10.

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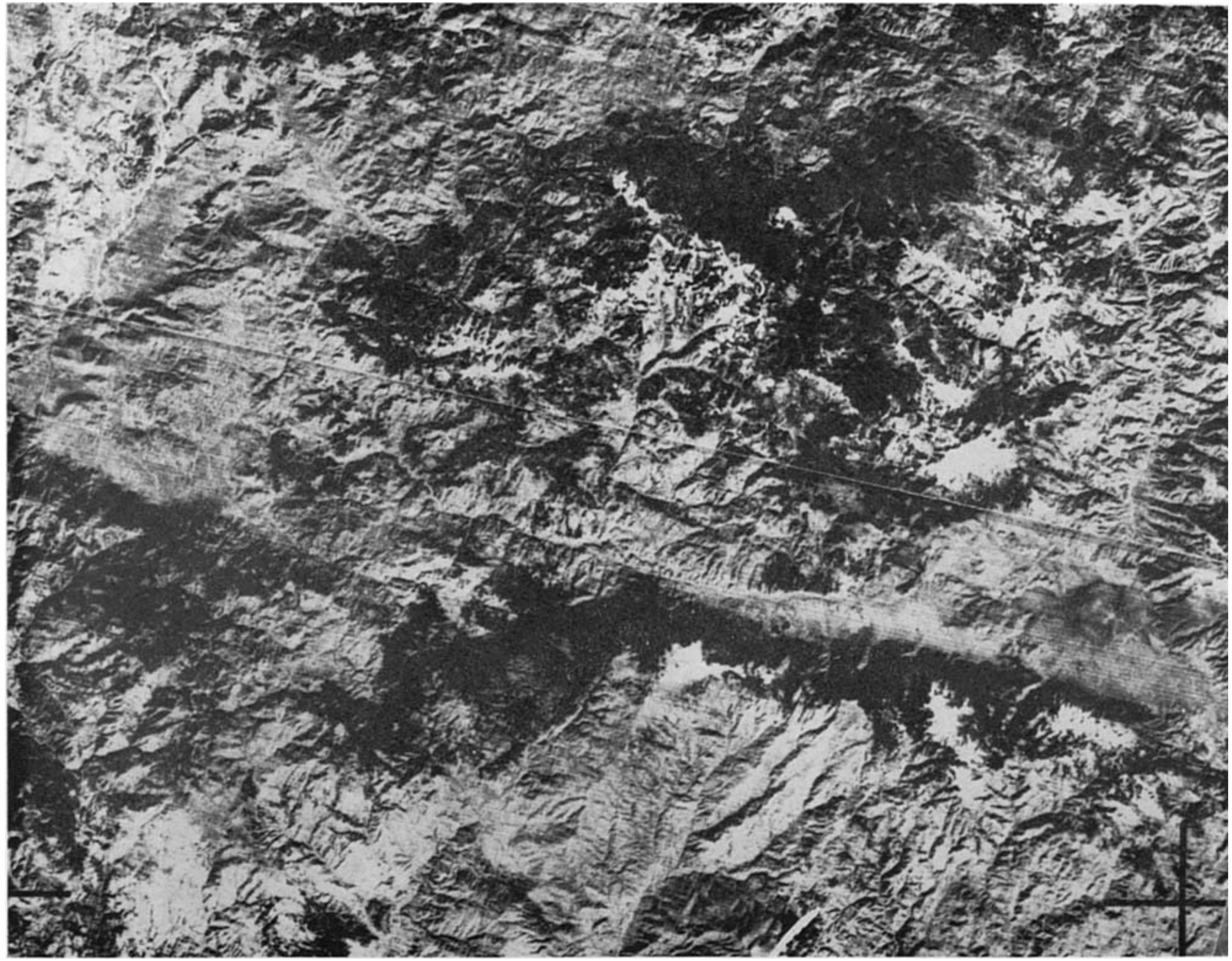


Plate 1. (b)

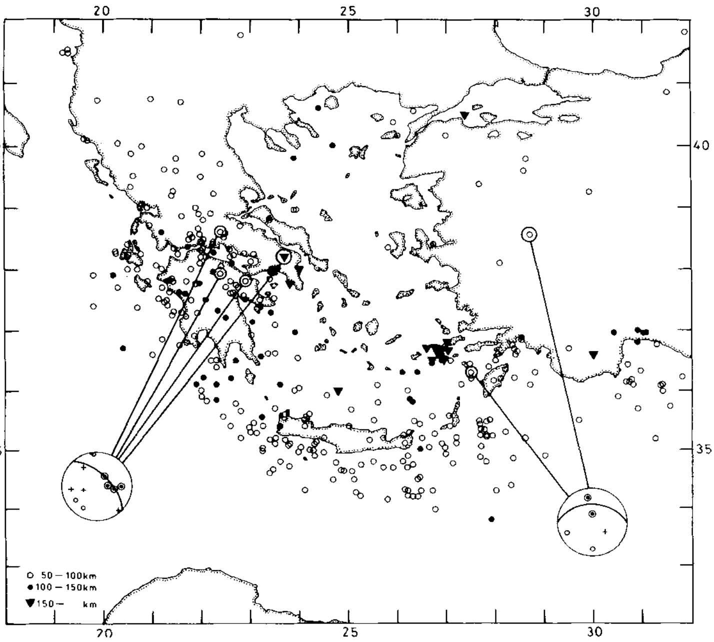


Figure 5. Epicentres of shocks below 50 km between 1961.1.1 and 1975.12.31 from the USCGS and NOAA. These depths are of variable accuracy for the reasons discussed in the text. The *P*, *T* and null axes of six intermediate shocks are shown in the small circles in an equal-area projection, connected to the epicentres concerned. The *T* axis is shown as a solid dot enclosed by a circle, the *P* axis by open circles and the null axis by crosses. The curved line shows the approximate dip and strike of the plane containing the earthquakes. Polarity observations are given in I or in Fig. 6.

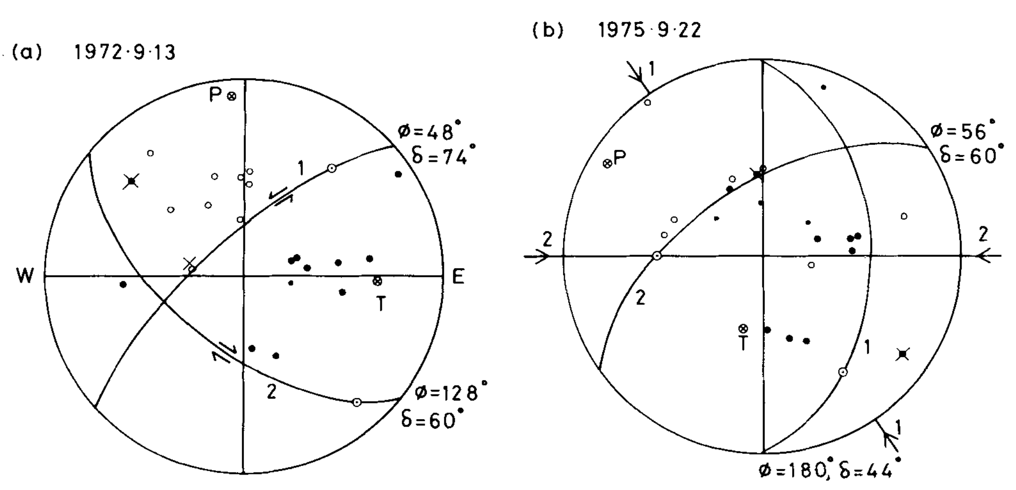


Figure 6. Polarity observations for two shocks whose depths exceeded 50 km. (a) 1972.9.13*P* axis, Az = 356°, dip = 9°; *T*, Az = 92°, dip = 33°; null, Az = 252°, dip = 55°. Plotted in Fig. 5. (b) 1975.9.22 *P* axis, Az = 302°, dip = 10°; *T*, Az = 195°, dip = 58°; null, Az = 36°, dip = 30°. Not plotted in Fig. 5, see text.

Crete, where little deformation is now occurring, must have influenced the direction of subduction along the Hellenic Arc. Until the history of the deformation behind the Arc is known in some detail it is not possible to make reconstructions to discover where pieces of the sinking slab were subducted.

Not all the intermediate events are related to present subduction. Five scattered events have been located beneath the northern Aegean which Papazachos (1973, 1976a) believes mark a sinking slab produced by a subduction zone in the northern Aegean. There is, however, no evidence of thrusting on the scale required (see Section 5) nor of associated vulcanism. If the depths are well determined these shocks presumably lie within material subducted at a trench which is no longer active. Several shocks deeper than 100 km have also been located beneath southwestern Turkey in a zone striking E–W. These may be related to a subduction zone further south marked by 1969.1.14 and 1975.4.30 in Fig. 3.

The *P*, *T* and null axes from six mechanisms of shocks below 70 km are shown in Fig. 5, in equal-area lower-hemisphere projections, together with a curve showing the approximate dip and strike of the active zone. One shock 1975.9.22 at a depth of 63 km is not plotted in either Fig. 3 or Fig. 5; it is not clear whether this event is intermediate or shallow, and its poorly-determined mechanism (Fig. 6) is unlike both groups. The event in Fig. 3 beneath western Turkey was located by the ISC at a depth of 65 km, whereas the USCGS obtained 72 km. It is unlikely that this event was shallow because its mechanism (I, fig. 23) is different from that of other events in the area which are known to be shallow (Fig. 8, Section 4).

The discussion above shows that the Hellenic Arc differs little from island arcs elsewhere which are subducting oceanic lithosphere. Whether the floor of the Mediterranean southwest of the Arc is oceanic is still unclear, since this question concerns not only the velocity structure, which is not yet well known and may have been changed from the typical oceanic velocity structure by metamorphism, but also its origin.

4 Western Turkey

The seismicity of western Turkey (Fig. 7) between 1961 and 1975 is concentrated in a number of bands running approximately E–W which correspond to the foreshock and aftershock sequences of major earthquakes. The three most recent sequences, those of the Gediz earthquake starting in 1969 March and of the shocks on 1971.5.12 and on 1975.3.27, were all associated with major normal faults, and fault breaks have been mapped for the first two of these sequences (Ambraseys & Tchalenko 1972; Arpat & Bingol 1969; Erinc *et al.* 1971). This activity is concentrated in a zone in western Turkey between 30½° E and 26° E, and the normal faults which extend both east and west of these limits appear to be much less active. The only mechanisms which are not normal faults are the strike slip events associated with the North Anatolian Fault (1967.7.22 and 1957.5.26), similar strike slip faults further west (1953.3.18, 1912.8.9), and a number of thrust events off the south coast which have already been discussed. Arpat & Bingol (1969) give a number of references to earlier workers who had recognized the normal faults in the field.

The dominance of the normal faults is also obvious in the satellite photographs and from the surface breaks of the larger earthquakes (Fig. 10) as Allen (1975) has pointed out. He shows *Landsat* photographs of the two areas marked 1 and 4 in Fig. 10, those of areas 2 and 3 are shown in Plate 1. The scarps of the normal faults show as linear segments joining at shallow angles, and separate the alluvial plains from the rougher terrain. Other normal faults are almost certainly active, cutting the alluvial plains and the hilly regions, but are not visible on the photographs. The faults on which the surface breaks of the Gediz earthquake

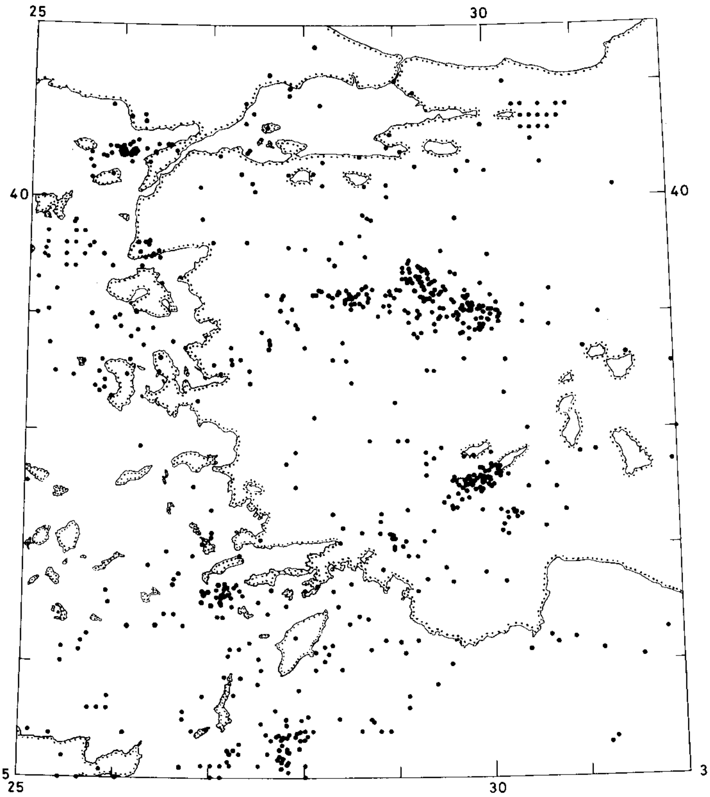


Figure 7. Epicentres of shocks with depths between 0 and 50 km in western Turkey, plotted on a conic projection which best fits the mozaic of *Landsat* images of the area. RMS error of fit is 2.7 km.

of 1970.3.28, mapped by Ambraseys & Tchalenko (1972), are a good example, and are not visible on the *Landsat* photographs. The other surface breaks in Fig. 10 are discussed in I except 1899.9.20, 1912.8.9, and 1971.5.12 taken from Ambraseys (1971), Allen (1975) and Erinc *et al.* (1971) respectively. Where fault plane solutions and surface breaks are available for the same shock the ambiguity between the fault and auxiliary plane can be resolved, and hence the slip vector determined. The azimuths of several slip vectors are shown in Fig. 10. Many other normal faults are also shown which have no recorded surface breaks, though otherwise they appear similar to those which have. These faults are visible on the *Landsat* photographs, and several are also shown by Crampin & Ucer (1975) and by Arpat & Bingol (1969). In the Aegean, however, the morphology and bathymetry are controlled by the faults only where they are active, because the sedimentation rate is high.

The motions in Fig. 10 show that the continent is being stretched in western Turkey. The motion occurs on many faults, probably many more than those which have been active since 1899. The slip vectors show a systematic change in trend. In the northeast of the region the fault plane solution and surface break of 1967.7.22 on the North Anatolian Fault are in good agreement and require a slip vector within a few degrees of 90° E. That of 1967.7.30 is less well controlled, since the strike would be changed if a crustal velocity was used. The slip vector of 1964.10.6 is well controlled by close stations, though the strike of the fault plane could be changed and remain compatible with the polarity observations. The slip vector is similar to that of 1953.3.18, which is also well controlled. The same is true of

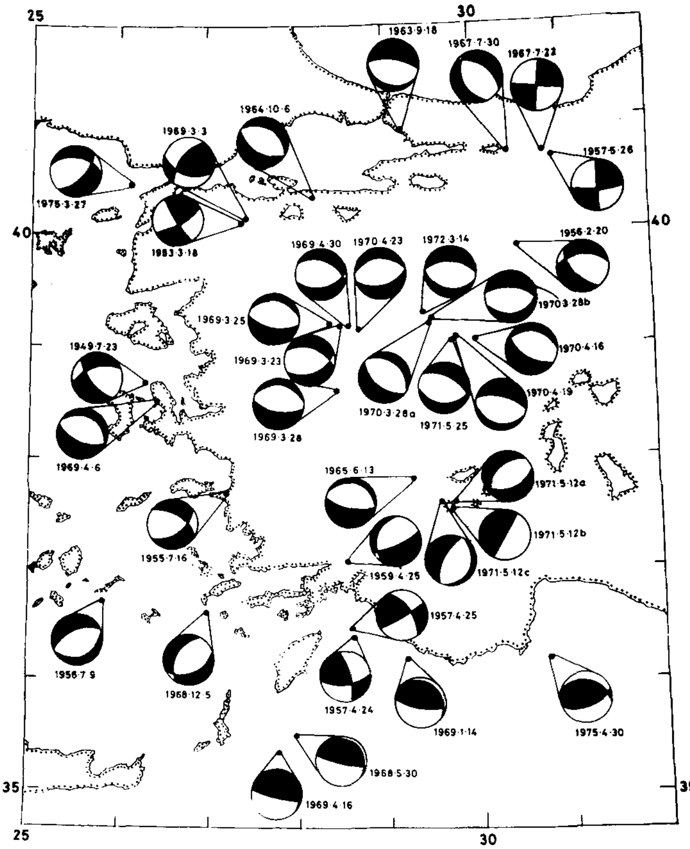


Figure 8. Focal mechanisms of shock in western Turkey. Projection the same as Fig. 6. Polarity observations in I and Fig. 9.

1970.3.28a. Both 1964.10.6 and 1970.3.28a had to be placed in the crust before two orthogonal planes could be fitted. The slip directions for 1969.3.23 and 1969.3.28 are less well constrained. The first shock was placed in the mantle in I; if it occurred in lower velocity rock the component of strike slip could be considerable. 1969.3.28 was placed in the crust, but the absence of compressional observations allows a strike slip component of motion if the crustal velocity is decreased. The auxiliary plane of 1971.5.12a is well constrained and the strike of both planes is insensitive to the crustal velocity. The slip vectors show a systematic change in trend, from ENE in the north to NW in the south. The motion therefore resembles the opening of a fan or spreading of a hand. Structures beneath the Aegean appear to have been produced by the same process, but the faults involved are mostly now less active (see Section 5).

Many sections across graben structures show the normal faults as planar structures with constant dip (Bott 1976; Arpat & Bingol 1969). There is, however, evidence that the dips of at least some of the normal faults in western Turkey become shallower at depth. The best evidence comes from 1970.3.28, whose fault break was mapped by Ambraseys & Tchalenko (1972). They measured the dip at the surface in several places to be between 60 and 70°. Where they could not measure the dips with confidence the outcrop of the breaks suggested these faults also were steeply dipping. The dip of the relevant plane of the fault plane

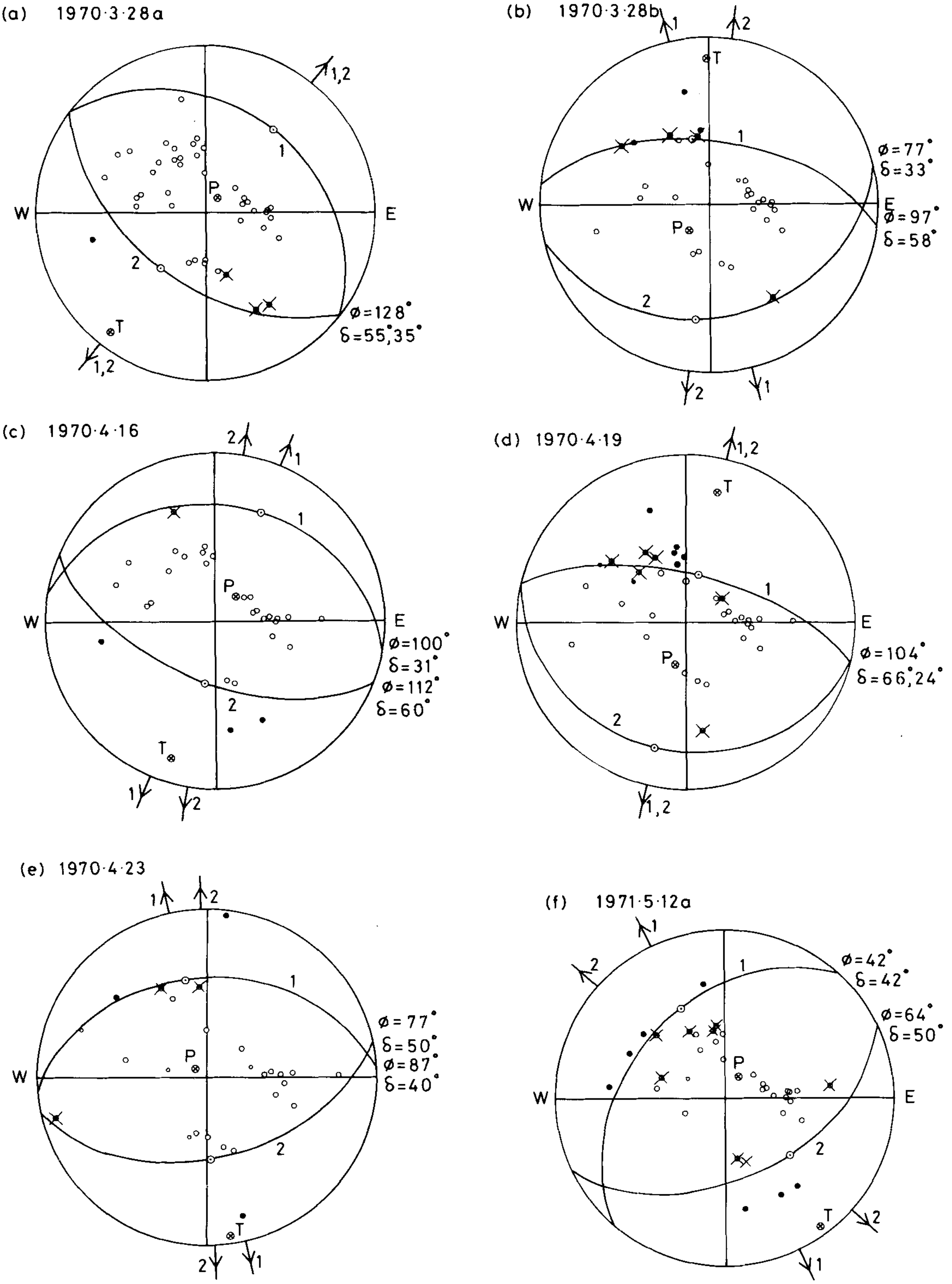
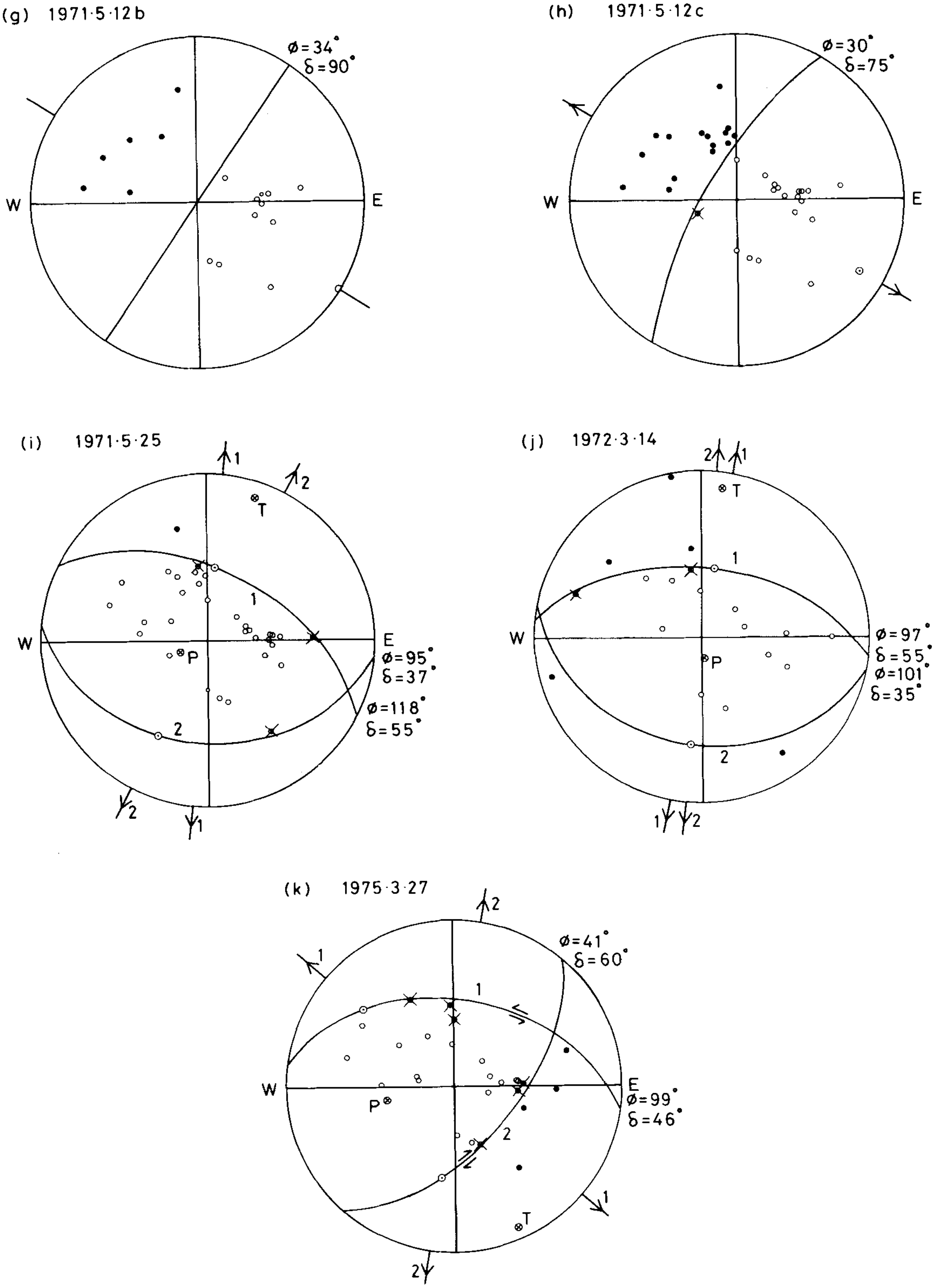


Figure 9. Polarity observations for mechanisms in Fig. 8; (e) and (j) placed in the mantle, the remainder in the crust with a *P*-wave velocity of 6.8 km/s.


(g) 1971·5·12b
ϕ=34°
δ=90°
W
E
(h) 1971·5·12c
ϕ=30°
δ=75°
W
E
(i) 1971·5·25
1
2
T
W
E
P
ϕ=95°
δ=37°
ϕ=118°
δ=55°
2
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(j) 1972·3·14
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1
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ϕ=97°
δ=55°
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2
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ϕ=41°
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2
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2

Figure 9. (g–k).

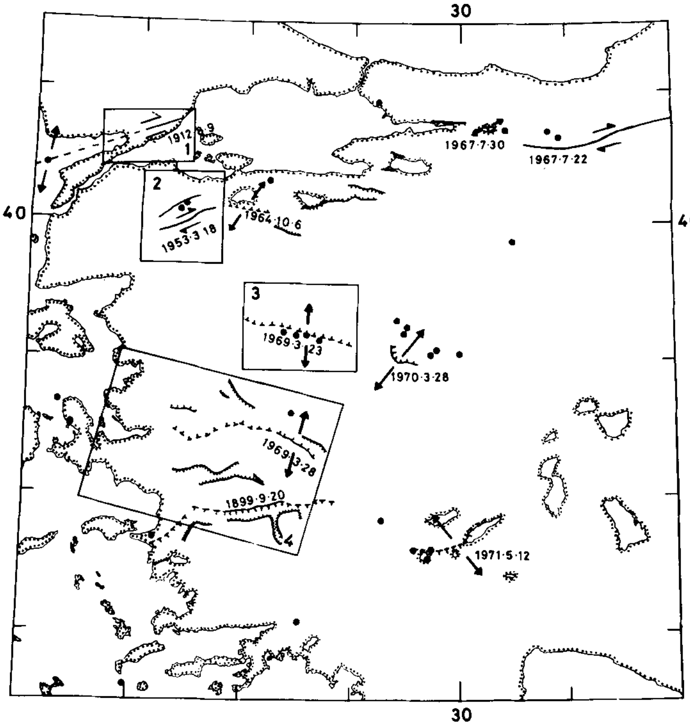

Figure 10. Surface breaks mapped after earthquakes are shown as heavy lines, with short lines on the downthrown side. Strike slip breaks are indicated as a single heavy line. Solid dots indicate the epicentres of the shocks in Fig. 8. Faults visible on the *Landsat* mozaic probably associated with shocks since 1960 are shown by heavy dashed lines. Fine lines indicate faults visible on the mozaic which are not believed to be associated with recent events but are probably active. The arrows show the direction of motion determined from the surface break and fault plane solution. The fault breaks are taken from: 1899.9.20 Allen (1975), 1912.8.9 Allen (1975) sense of motion not determined, 1953.3.18 Ketin & Roesli (1953), 1964.10.6 Ketin (1966), 1967.7.22 Ambraseys & Zatopek (1969), 1967.7.30 Ambraseys & Zatopek (1969), 1969.3.28 Arpat & Bingol (1969), 1970.3.28 Ambraseys & Tchalenko (1972), 1971.5.12 Erinc *et al.* (1971). The *Landsat* images of regions 1 and 4 correspond to Allen's (1975) figs 12 and 13 respectively, and regions 2 and 3 are shown in Plate 1. Projection that of Fig. 7.

solution is 35° N, and the dip but not the strike of the plane is well controlled. Though the dip depends on the source velocity used, if the velocity is decreased the auxiliary plane must become steeper. Since the fault plane is at right angles to the auxiliary plane its dip becomes shallower. It does not, therefore, appear possible to force the surface observations to agree with the fault plane solution. The evidence for the other two shocks is less good. The surface break for 1969.3.28 was mapped by Arpat & Bingol (1969) but they do not give any dips. However, their fig. 1 shows a number of faults with steep inclinations, whereas the fault plane solution shows the active plane had a dip of 29°. The third example is 1964.10.6. A surface break was mapped by Ketin (1966) and probably was produced by this shock. If so, then the dip of the fault plane is 36° determined from the fault plane solution. The epicentres of all three shocks are displaced to the north of their surface breaks, but in the absence of accurate focal depths the dips of the fault planes cannot be estimated. Furthermore, the epicentres of two strike slip shocks, 1953.3.18 and 1967.7.22, are also displaced northwards and suggest that this behaviour may be due to systematic errors.

5 The Aegean Sea

In the last five years marine geophysicists have surveyed the Aegean in some detail and an excellent bathymetric map has been published (Morelli, Pisani & Gantar 1975). The principal bathymetric features were pointed out by Maley & Johnson (1971) and are visible in Figs 11 and 12. The graben systems of western Turkey extend beneath the Aegean as troughs, though the southern ones are relatively inactive. The NE trend of the structures beneath the Aegean north of 38° N is truncated by NNW troughs NE of Greece and Evvia. Most marine geophysicists have argued that the structures beneath the Aegean have been produced by extension (Jongsma 1975; Jongsma *et al.* 1977; Needham *et al.* 1973), though Maley & Johnson (1971) were undecided whether the Anataolian Trough in the northern Aegean was a rift or a trench.

The fault plane solutions in Fig. 12 strongly support the existence of extension. Two new solutions are for events on 1956.7.9 and 1975.3.27. The first of these shocks was shown as dominantly strike slip in I, though many of the short-period observations were inconsistent with the focal planes chosen. However, Shirokova (1972, fig. 59) has published a more consistent mechanism for this event which also agrees better with those of other shocks in the

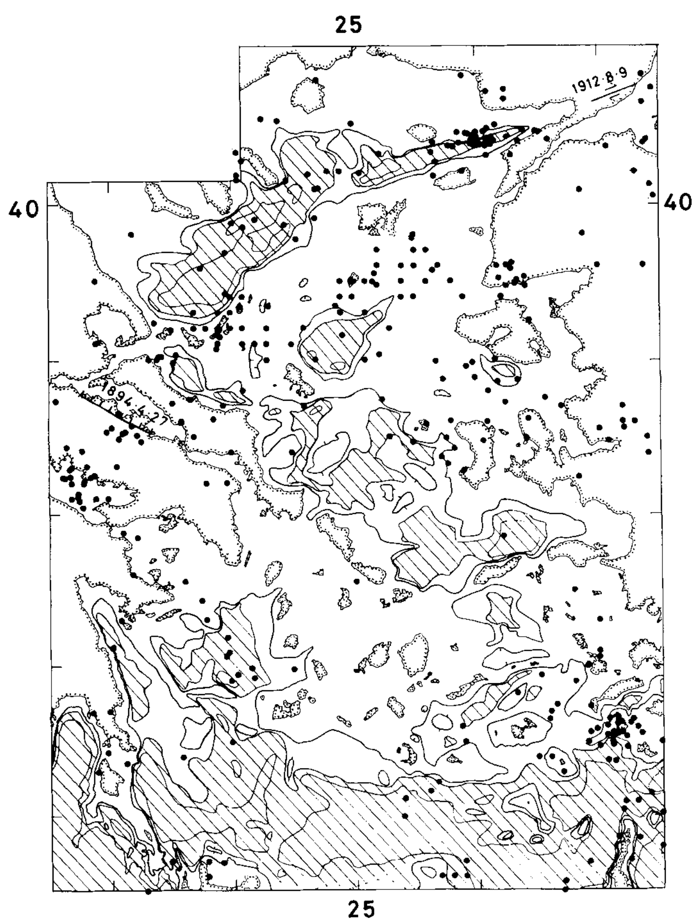

Figure 11. Distribution of epicentres of events shallower than 50 km. Contours are at depths of 400, 600 and 1000 m taken from Morelli, Gantar & Pisani (1975), shaded below 600 m.

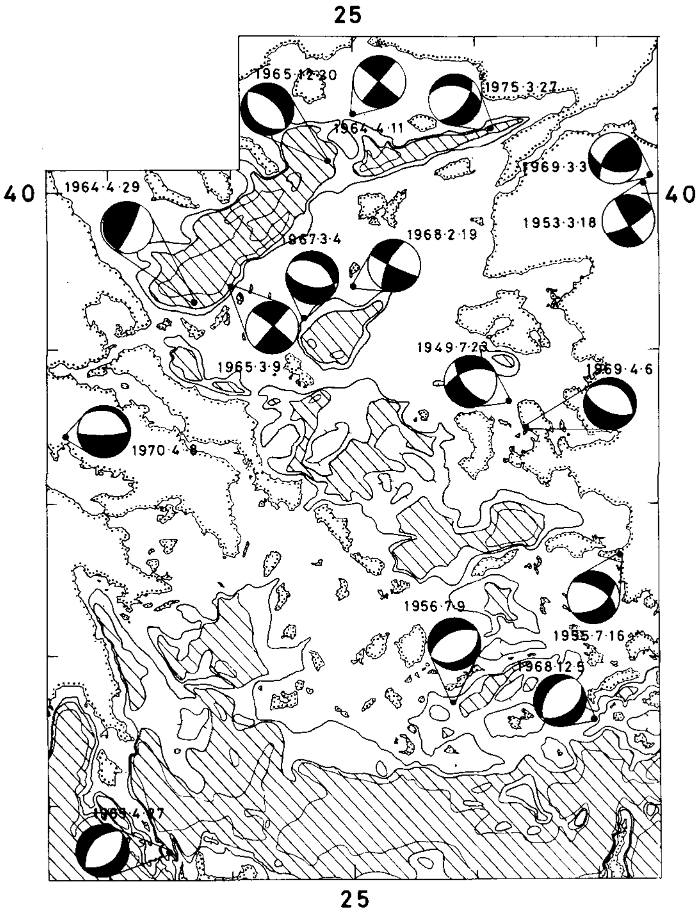

Figure 12. Focal mechanisms of shallow shocks beneath the Aegean. Polarities in Figs 9, 16 and I.

neighbourhood, and is therefore to be preferred. The other new mechanism, for 1975.3.27, is shown in Fig. 9(k) and is very well constrained to be a normal fault with a considerable component of strike slip motion. Both Papazachos (1976a) and Ritsema (1974) show many mechanisms for shocks in the northern Aegean which are thrusts but do not show the polarity observations. The only thrust in Fig. 12, which is believed to contain all the reliable mechanisms, occurs at the SW end of the Anatolian Trough where the change in trend occurs. In the same area Jongsma (1975) believed he could see thrusts on reflection records.

The mechanism of 1975.3.27 together with Allen's (1975) description of the surface break of 1912.8.9 are consistent with a strong right lateral strike slip motion on the eastern part of the Anatolian Trough. This part also has a negative gravity anomaly of -50 mgal (Needham *et al.* 1973), which is a typical feature of oceanic fracture zones. There is some evidence that the western part of the Anatolian Trough contains a larger component of opening. The (rather poor) solution for 1965.12.20 contains a large component of normal faulting. Both focal planes are parallel to features described by Needham *et al.* (1973) which are normal to the trend of the trough. Two substantial shocks, 1967.3.4 and 1968.2.19, have occurred south of the Anatolian Trough and are not associated with any major bathymetric feature. There is, however, a zone of disturbed sediment in this area clearly visible on reflection seismic records (Jongsma 1975). The aftershocks of 1968.2.19 have recently been relocated by North (1977) who found that they lay on a NE–SW trend. If this trend follows the fault plane then the motion must be right lateral, as had previously been supposed in I. Recently Lalechos & Savoyat (1977) have obtained deep penetration seismic reflection records covering the area of the Anatolian Trough and the shallow regions further north. They found evidence of widespread normal faulting on two trends, one parallel to the Anatolian Trough, the other striking between NW and NNW. They found no evidence of shortening in the area.

In the southern part of the Aegean there is little seismic activity, and all the mechanisms show normal faulting. Those in the east, 1956.7.9 and 1968.12.5 have nodal planes striking parallel to local troughs, which are probably grabens bounded by normal faults. Their trends are parallel to those on land and the direction of opening is the same. The solution NW of Crete is more surprising. Both nodal planes are approximately normal to the Hellenic Arc, though the strike of only one of them (plane 1 in I, fig. 21(m)) is well controlled. Similar normal faults striking N are known on Crete (Angelier 1977), and have also given rise to the Gulf of Corinth and other parallel graben in Greece (Section 6).

There has been considerable speculation about whether the North Anatolian Fault extends beneath the Aegean. The information in Figs 11 to 13 suggest that it does not and that the motions are taken up on several structures with considerable components of normal faulting. This process starts with 1967.7.30, and the bathymetric map of the Sea of Marmara (Fig. 13) suggests it is a graben. The opening on this graben is probably oblique and parallel to the slip vector for 1967.7.30. Furthermore this slip vector is parallel to the trace of the strike slip fault which broke in 1912.8.9 (Allen 1975). The southwest extension of this fault beneath the sea is marked by a narrow trough (Fig. 13), which is slightly oblique to the eastern part of the Anatolian Trough and forms its eastern boundary. The shock on 1975.3.27 has a considerable component of strike slip motion, which is right-handed if plane 2 is the fault plane. The strike of plane 1, and hence of the slip vector, depends strongly on the velocity of the origin, and has a larger east–west component than that shown if the velocity exceeds 6.8 km/s. These fault plane solutions and maps of surface breaks suggest that the motion is taken up to the west on a number of approximately E–W grabens. The fault plane solutions show that the N–S component of the motion increases

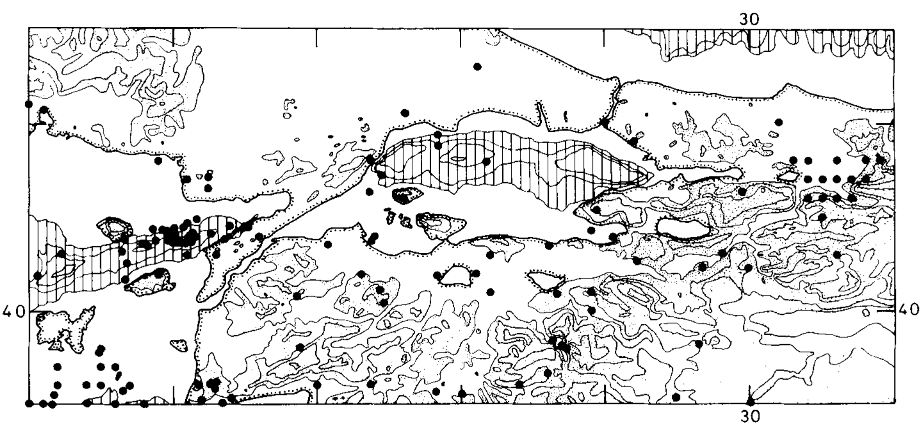

Figure 13. Bathymetry and topography in the neighbourhood of the Sea of Marmara, taken from US Navy Map No. 5622 (1st edn). Submarine contours 100 (200) 700 fthm, except in the Black Sea where only the 100 and 500 fthm are shown. Subaerial contours 1000, 2000, 3000, 5000, 7000 ft. Focal mechanisms and surface breaks in Figs 8 and 10.

(a)

Plate 2. *Landsat* images of regions 1 (a), 2 (b) and 3 (c), in Fig. 17. (a) 811 020 848 55G, 811 020 848 25G (b) 811 040 859 55A, 811 750 854 15A, 811 020 848 55G, 811 020 849 15G, (c) 811 020 849 15G, 811 920 849 45N.

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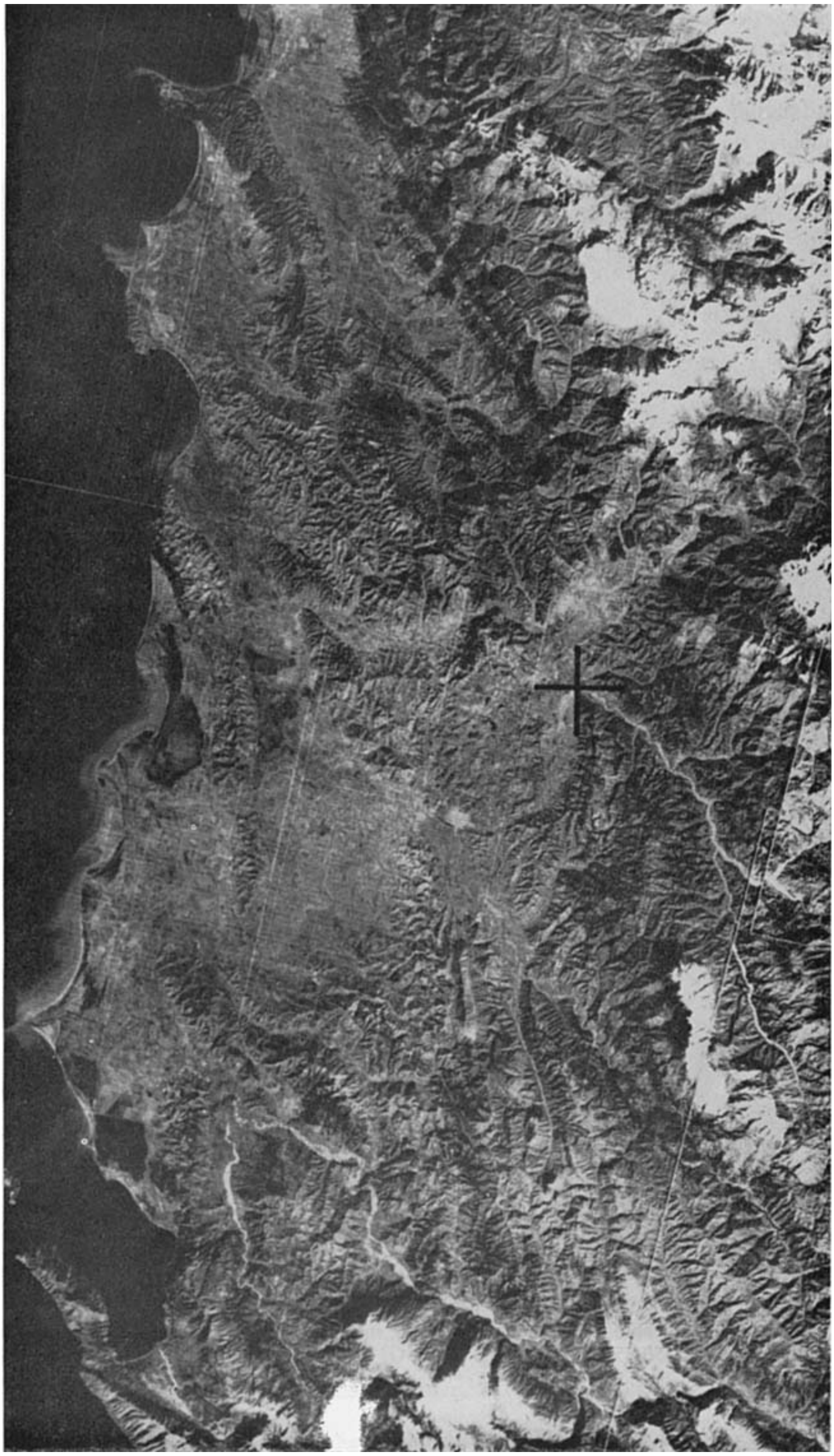

Plate 2. (b)

Plate 2. (c)

westward. This increase can account for the change in the trend of the strike slip faults from E–W for 1967.7.22 on the North Anatolian Fault to ENE for 1912.8.9 and 1953.3.18 to NE for 1965.3.9 and 1968.2.19.

6 Greece, Albania, Yugoslavia and Bulgaria

This region is less well understood than western Turkey, and little attempt was made in I to describe the tectonics of the region. Principally because of the *Landsat* photographs it is now possible to discuss the area in more detail and to give a general account of the deformation. As already mentioned (Section 1) there is no evidence to support the suggestion made in I that the Anatolian Trough is connected to the Gulf of Corinth by a NE trending strike slip fault.

The distribution of seismicity (Fig. 14) shows that the activity associated with the Hellenic Arc bends eastward north of the island of Levkas and north of this point is almost entirely confined to the land. Only in the extreme NW, where the Albania–Yugoslavia border reaches the Adriatic, are off-shore structures again seismically active. This behaviour is probably caused by the existence of the rigid Apulian block beneath the sea to the west, which is stronger than the heavily deformed belts of Albania, Yugoslavia and Greece. The other major active zone follows the Gulf of Corinth eastward towards the Aegean. The part

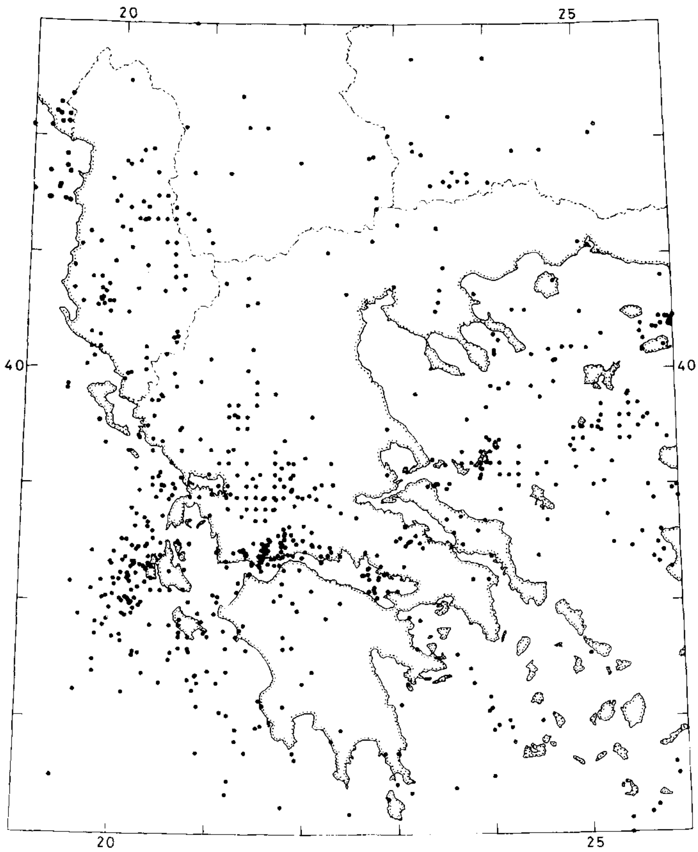

Figure 14. Epicentres of shocks with depths between 0 and 50 km, plotted on a conic projection which best fits the mozaic of *Landsat* images of the area. RMS error of fit is 7.3 km.

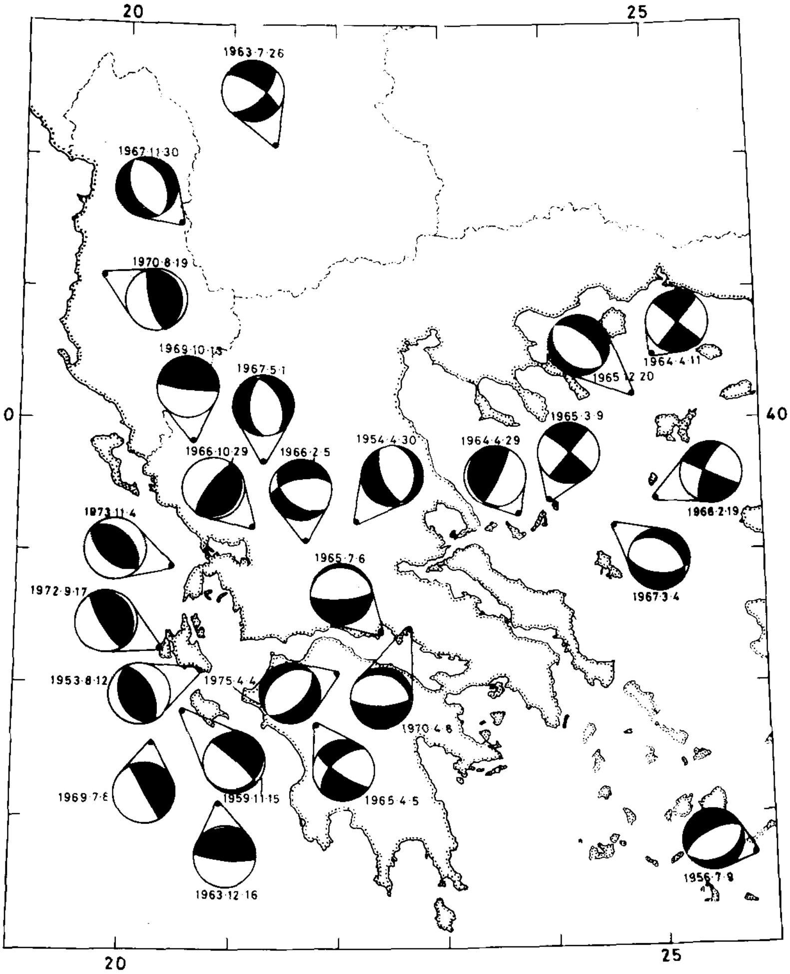

Figure 15. Focal mechanisms of shallow shocks in the area covered by Fig. 14 and in the same projection. Polarity observations in I and Fig. 16. Crampin & Ucer (1975) pointed out that 1969.10.13 was misplotted in western Turkey in I.

of Greece south of the Gulf has some active normal faults (Dufaure 1965) but their displacements are considerably smaller than that on those further north along the southern margin of the Gulf of Corinth.

The fault plane solutions in Fig. 15 show that, except in western Greece and Albania, most of the deformation is produced by normal faulting. This observation agrees excellently with the field mapping of Mercier *et al.* (1976) in the Gulf of Corinth region and further north. They found Kefallinia had been deformed by thrusting during most of the Pleistocene whereas further east normal faulting had been active. They also showed that the deformation regime was compressional in the earliest Pleistocene. Their observations suggest that the boundary separating normal faulting from thrusting can change position rather rapidly. The division of the active region north of Kefallinia into a western thrust belt and an eastern one of normal faulting is very obvious on the *Landsat* photographs. Several normal faults are clearly visible (Fig. 17). Many others probably exist but are not so easily visible either because they are small or because the Sun elevation is too great. A portion of the photmosaic in Plate 2(a) shows one of these faults. It is more linear than those in Western Turkey,

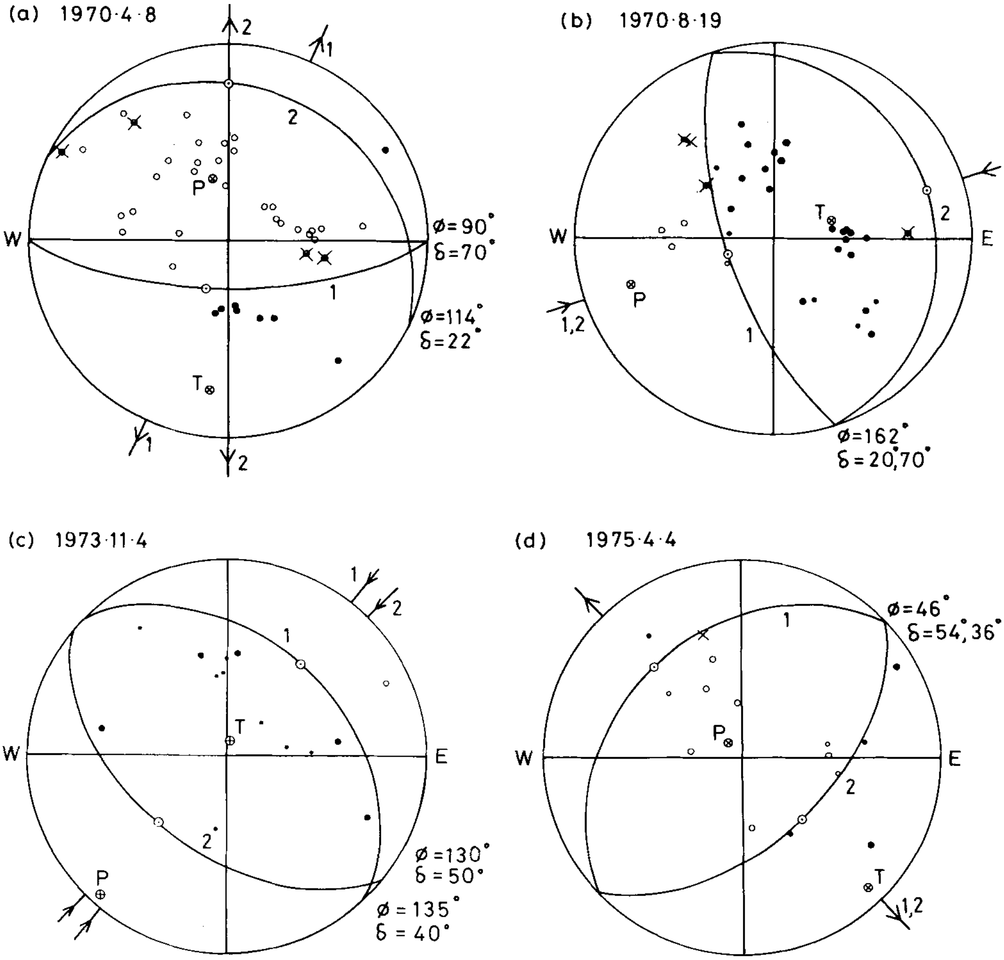

Figure 16. Polarity observations for mechanisms in Fig. 15. (b) in the crust, the remainder in the mantle.

probably because it follows the trend of older geological structures. A number of normal faults have also been mapped by Arsovsky & Hadzievsky (1970), who show the arcuate structure in Plate 2(a) as a fault. This structure marks the northern extension of a number of major normal faults. Therefore if it is a fault the motion on it should be right-handed strike slip in a direction approximately parallel to one of the planes of the fault plane solution for 1963.7.26 nearby (plane 2 in I, fig. 20(e)). This choice is consistent with the regional tectonics (see I and Section 7) but not with the field observations of surface breakage (see below).

The thrust zone to the west looks very different (Plate 2b, c) and consists of a number of large amplitude open folds. The style of deformation is also found in the northeast Caucasus, southeast Turkey and the Zagros mountains in Iran. In all these regions the deformation is produced by thrusting and the succession contains mobile salt horizons. A region of Fig. 17 north of the Gulf of Corinth has been mapped in detail and the subsurface explored by drilling (British Petroleum 1971). That study showed that a number of the thrust faults outcropped in evaporite horizons, and that major thicknesses of halite existed at depth. It is therefore not surprising that the faults are not visible on the *Landsat* mosaics. Further

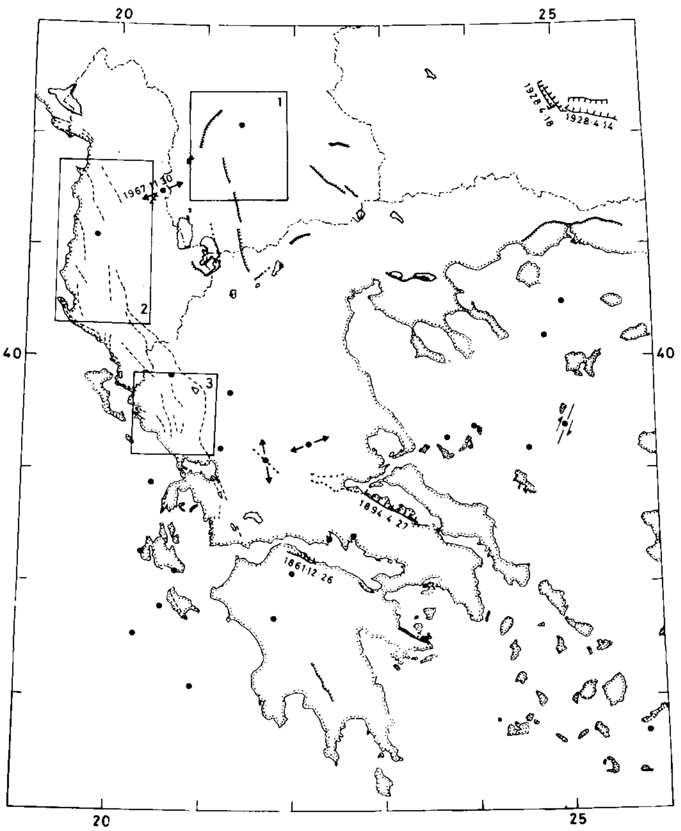

Figure 17. Surface breaks and faults visible on the *Landsat* images (see Fig. 10 for details). Projection that of Fig. 14. The fault breaks are taken from 1861.12.26 Richter (1958), 1894.4.27 Richter (1958), 1928.4.14 and 1928.4.18 Richter (1958), 1967.11.30 Sulstarova & Kociaj (1969) and Ambraseys (private communication). The *Landsat* images of areas 1 to 3 are shown in Plate 2.

north IFP (1966) have carried out a similar detailed investigation in approximately the region covered by Plate 2(c). They find widespread evidence of thrusting and of decoupling caused by evaporite horizons. Both investigations are more concerned with the geometry of the folds and faults and their relationship to the stratigraphy than whether the deformation is continuing. However both here and elsewhere the correlation between open folds which dominate the geomorphology and active thrusting is striking. This correlation suggests that the area of western Greece, Albania and Yugoslavia where such folding is obvious on the satellite mosaics (Fig. 17) is associated with thrusting and salt tectonics. The thrusting region may be more extensive than that shown.

Several earthquakes have been associated with surface breaks, all but one showing normal faulting. The most impressive are the breaks associated with 1928.4.14 and 1928.4.18 in southern Bulgaria (Fig. 17). They were mapped by Jankof (see Richter 1958) and follow the two sides of a graben. These faults are not visible on the *Landsat* mosaics. Though no fault plane solution is available for either shock, the absence of appreciable strike slip motion on the E–W part of the graben system suggests N–S extension. The other major surface break in the region was associated with a shock on 1894.4.27 in eastern Greece. This break is also described by Richter (1958), who again remarks that there is little evidence of strike slip motion. He also describes that of a smaller shock on 1861.12.26 on the southern shore of the Gulf of Corinth. Though he attributes the break to secondary slumping, the sense and form of the motion agree with that expected in the area (see also Section 7). The only other surface break which has been described in any detail was associated with 1967.11.30 (Arsovsky 1970; Sulstarova & Kociaj 1969). The fault trended about 30° E and was downthrown on the SE side. Though this trend is in poor agreement with the fault plane solution in I, the strike of the nodal plane labelled 1 in I, fig. 20(r) is poorly constrained. The polarity observations themselves are not inconsistent with the mapped break. This uncertainty results from the absence of any close WWSSN stations in the NE quadrant covering the USSR and other eastern European countries. The auxiliary plane 2, and hence the slip direction, is better controlled, but the strike slip motion is then left lateral, which was not observed. Ambraseys (1975) lists three other shocks with possible surface breaks for which fault plane solutions are shown in Fig. 15. The strike of the break from 1954.4.30 is consistent with the polarity observations in I, fig. 20(c) and suggests plane 1 is the fault plane. That for 1963.7.26 agrees well with the sense of motion and strike of plane 1, though the surface break was not well defined. The last one, for 1966.10.29, is not compatible with the polarity observations. Either the fault involved has a complicated shape, or the break was not associated with the earthquake. Therefore the slip direction cannot be determined. The ambiguity between the fault and the auxiliary plane has been resolved for one shock in the region with no known surface break. Fitch & Muirhead (1974) used the aftershocks of 1966.2.5 to show that the plane dipping SW was the fault plane, hence the extension was approximately NS.

The slip vectors obtained from the fault plane solutions are plotted in Fig. 17, though none of them are well constrained. Widespread normal faulting on faults with NW trends is visible on the satellite photographs, and is clearly shown by drilling in the northwestern Aegean (Lalechos & Savoyat 1977). This direction is parallel to that of a few small shocks beneath the southwestern part of the Aegean, where Jongsma *et al.* (1977) have demonstrated normal faulting still continues. Near the Gulf of Corinth the extension direction is more N–S. The strike of many of the normal faults and some of the thrusts is probably inherited from the older geological structures (IFP 1966).

The most obvious problem in accounting for the observed deformation is the relationship between the thrusting and normal faulting in the northwestern part of the area. Since the thrust and normal faults have similar strikes and dips there is almost no overlap of the dilatational quadrants of the fault plane solutions. Hence it is not possible to produce the observed motions with a single stress field even if both sets of faults are reactivated (McKenzie 1969b). That the stress system changes so completely is surprising and not easily explained (see Section 9).

7 Geometry and mass conservation

This section and the two which follow are concerned with kinematics, thermal structure and dynamics of the Aegean region. Some of these questions can be examined in considerable detail because so much more is known about the Aegean region that about other similar structures which are no longer actively deforming.

Fig. 18 shows a summary of the present motions and directions, and the major difference between it and Fig. 1 is in the structures and motions shown in northern Greece, Albania and Yugoslavia. To a limited extent plate tectonics provides a useful description of these motions. Where the motion is taken up on a major structure such as the North Anatolian Fault or the Hellenic Arc, and where the regions on both sides are undergoing little deformation, the simple ideas of plate tectonics can successfully describe the observed slip direc-

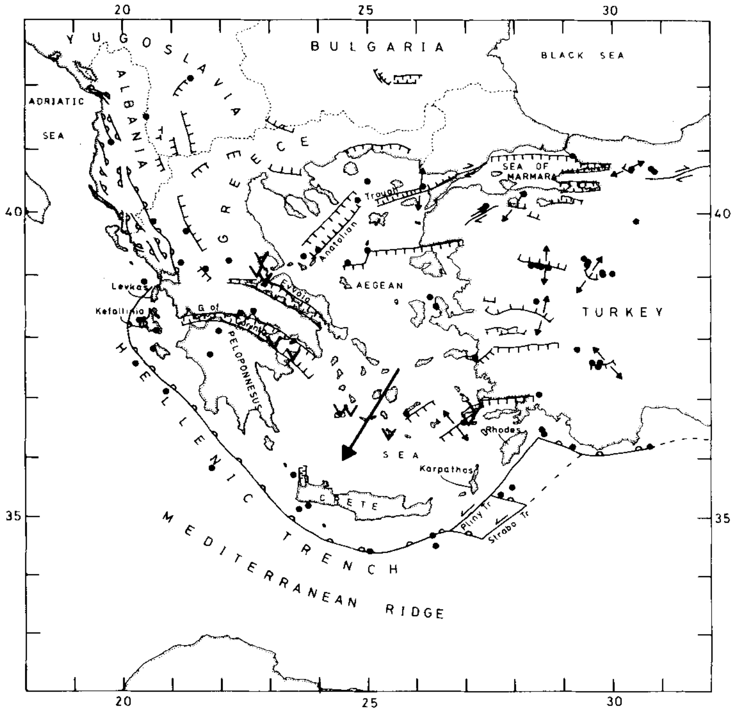

Figure 18. Summary of the present deformation of the Aegean region. Normal faults are shown as long curved lines with short lines at right angles on the downthrown side. Thrust faults are lines with open semicircles on one side. Except for the thrust fault along the SW side of the Hellenic Arc, where the semicircles are on the overriding side of the fault, the side on which the semicircles are has no significance. Solid dots mark epicentres of shocks for which mechanisms are available and which are believed to be produced by the observed surface deformation. The directions of motion obtained from fault plane solutions are shown by arrows. The long heavy arrow shows the direction of relative motion between the Aegean and Africa taken up by the Hellenic Arc. Heavy Vs mark sites of recent vulcanism, taken from Ninkovich & Hayes (1972). Because the period for which accurate instrumental locations for epicentres is so short, and because faults not visible on the *Landsat* mozaics have produced at least two major shocks in this area since 1928, this figure should not be used as a guide to seismic risk. Many more faults than those shown must be active.

tions. But the theory is of little value in regions such as northern Greece or western Turkey where the deformation is spread over a zone. Within these zones the motions are not random, and the type of faulting and the slip vectors vary systematically with position, but in a way which is not simply related to the relative motion of the aseismic regions on either side of the deformation zone. For example, in western Turkey the slip vectors may vary systematically from almost E–W in the north to NW–SE in the south. This behaviour could be produced by the rotation pole between the Aegean and Turkey being in central Turkey, but there is no evidence of corresponding behaviour in the thrust mechanisms along the Hellenic Arc. In northwestern Greece the behaviour is more surprising, since there appears to be no aseismic region between the zones undergoing extension and shortening. Dewey & Sengor (1978) have also remarked that plate tectonics is not useful in the Aegean area where the normal faulting is not confined to a narrow zone. It is important to stress that these observations do not imply any failure of the concepts of plate tectonics. There is no evidence that the total slip across the active zones in the Aegean is different from the relative motion of the aseismic regions. The problem is that the relative motion does not predict the details of the deformation within the zones and hence it is not useful in such regions. Why the deformation zone is so wide, especially in western Turkey, is still unclear. The most obvious explanation is that the pre-existing geological structures are EW and cannot easily take up a N–S shear. This suggestion was made in I but such faults have not been found by field mapping (Brinkmann 1976; Dewey & Sengor 1978).

In I the velocity of the Aegean plate relative to Africa and Eurasia was estimated from the difference between the azimuth of the slip vectors in the northern part of the Aegean and along the Hellenic Arc. Since the difference between the two angles was small it then seemed unlikely that the velocity estimates were reliable. If the new observations are included the mean azimuth in the northern Aegean is 215° and that on the Hellenic Arc is 211°. The difference is not significant, and these results justify earlier doubts. The only other method of estimating the velocity is from the relative motion between Turkey and Eurasia, given in I as 4 cm/yr in an E–W direction in western Turkey. The slip vectors between the Aegean and Turkish plates are N–S, and this motion requires a velocity of about 7 cm/yr for the Aegean in the direction of the arrow in Fig. 18. Because this estimate depends on the details of the tectonics in eastern Turkey it could be reduced by a factor of about two. Much larger reductions cause the slip direction between the Aegean and Eurasian plates to differ significantly from that between the Aegean and African plates, and would therefore disagree with the seismic observations.

Numerous authors have described the extension in the Aegean region as back-arc spreading. This concept was originally rather precisely defined by Karig (1971) to be extension occurring behind island arcs along the line marked by the volcanoes, but is now often used for any sort of spreading behind island arcs. The extension in Fig. 18 clearly did not start along the volcanic arc, and extends both E and W of the subduction zone. Therefore, whether it is described as back-arc spreading depends on which definition is used. Such a description is useful only if there is a relationship between the extension and the subduction, and this is in general far from clear.

The geology of Greece, Turkey and the Aegean Islands suggests that they all once formed parts of the same orogenic zone (Aubouin 1973; Bernoulli, de Graciansky & Monod 1974; Aubouin *et al.* 1976; Brunn *et al.* 1976). However, the present elevation differences between Greece and Turkey on one hand and the Aegean on the other are substantial. The difference in elevation corresponds to a similar difference in crustal thickness. Makris (1975) and Makris & Vees (1977) have shown that the crustal thickness beneath Greece and Turkey is between 40 and 50 km, whereas beneath the Aegean Sea the crustal thickness varies between 22 and 32 km (Jongsma *et al.* 1977; Makris 1976; Makris & Vees 1977). These thicknesses are based on gravity and seismic refraction observations. Woodside (1975, and private communication) has also estimated similar crustal thicknesses, though his seismic control is less extensive than that of the other authors. The simplest explanation of these observations is that the thin crust beneath the Aegean has been produced by stretching the orogenic belt by about a factor of two. The sea-floor shows evidence of such stretching in the past in the southern Aegean, where it is still continuing (Jongsma *et al.* 1977), though much less vigorously than in western Turkey and Greece. Fault plane solutions and surface breaks show that normal faulting is vigorous in regions of western Turkey, Bulgaria, Greece and Albania where it is not visible on the *Landsat* photographs. Both these observations can be explained if the faulting now extends over a larger region than it did in the Pliocene, and suggest that the region undergoing strong deformation by normal faulting has progressed outwards from the southern Aegean. The motions in Fig. 18 suggest that the westward movement of Turkish plate carries western Turkey through an extensional zone where the continent is stretched and its thickness halved. Afterwards, little further deformation occurs. This process doubles the surface area in the deforming zone without generating any new crust, and is therefore quite unlike the process at an oceanic ridge.

Cross-sections of normally faulted regions often show faults with constant dips cutting through the crust. However, many normal faults in regions which have been carefully mapped are known to decrease with depth and are called listric faults. Seismic reflection profiles across active faults in the Basin and Range province in the western United States clearly show a decrease in dip with depth (Snelson, private communication). Similar faults have been mapped in the area (Young 1960; Longwell 1945). In the northwestern Aegean seismic reflection profiles show that listric faults are widespread here also (Lalechos & Savoyat 1977), and have been mapped on land in Evvia (Mercier, private communication). The fault plane solutions discussed in Section 4 also suggest that the dip decreases with depth. Murawski (1976) has used a similar argument in the Rhine Graben to argue that the extension there is also occurring on listric faults. Such faults are probably more common than is usually believed, and can account for the gravity field observed above graben structures (Thompson 1959) without complicated return flows.

The geometry of listric faulting is shown in Fig. 19. At the surface break the dips are steep but the decrease with depth causes back tilting of the subsided blocks. When displacement occurs more subsidence takes place where the fault is steep and hence the thickness of the material filling the graben increases towards the fault. For the same reason any river or lake in an active graben is displaced towards the faulted side. Unlike thrust and strike slip faults, where gravitational forces oppose the motion or are not involved, movement on normal faults releases gravitational energy. It is therefore possible to have a continuous transition from high stress normal faulting, which mostly releases elastic energy, to slumping, which principally releases gravitational energy. It is therefore not surprising that the shapes of seismically active listric faults and growth faults are similar. An important result of the geometry in Fig. 19 is to cause estimates of extension made from the dip and throw of faults at the surface to be too small, quite possibly by as much as an order of magnitude. Such errors could explain the discrepancy between the geological and geophysical estimates of the extension in the Gulf of Suez and East Africa. The thermal structure is also very strongly affected by the magnitude of the stretching (see Section 8), and hence by the fault geometry.

The discussion above is concerned with the brittle part of the crust. There need be no close relation between the deformation in this part and that in the hotter part below where the deformation is aseismic and occurs through creep. The style of deformation observed in

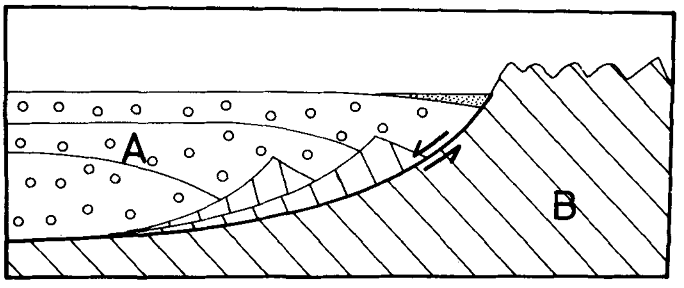

Figure 19. Sketch to show the relationship between the surface breaks of normal faults and the deeper motions (heavy line) with shallower dips. A, alluvium; B, basement. The fine lines within the alluvium mark former positions of the alluvial surface. The geometry requires internal deformation of the downthrown material overlying the fault, and produces downwarping near the surface break, often leading to the formation of a lake, shown with fine dots.

high-grade metamorphic terranes is very different from that observed nearer the surface and the relationship between the two is not yet understood.

The other major mass conservation problem is the fate of the sediments swept into the Hellenic Arc. The present crustal thickness in the eastern Mediterranean is at least 20 km (Makris 1978). If it consists of oceanic crust overlain by sediment then the sedimentary thickness is about 15 km. The depth of the intermediate shocks beneath the Aegean is at least 150 km, hence about 200 km of plate must have been consumed. There is no obvious place to put this amount of sediment, nor is it easy to understand how it could be subducted.

8 Thermal structure

It has been widely believed for ten years that the heat flow behind island arcs is anomalously high, and a number of theoretical models have been proposed to explain the observations, such as shear stress heating and erosion of the base of the lithosphere. Recently, however, the importance of hydrothermal circulation through the oceanic crust has been recognized as an important heat transfer process. Heat transport by convection across the water–rock boundary does not contribute to the temperature gradient at the interface, hence the conductive heat transport is less than the total. Sclater, Crowe & Anderson (1976a) have collected from the deep oceans those observations which they believe are reliable and find that the mean heat flow as a function of age is considerably greater than the mean of all observations. Also the reliable curves for heat flow and for depth as a function of age both agree with the curve calculated for a cooling plate model. Furthermore those marginal basins whose age is known have heat fluxes which lie on the same curve for heat flow as a function of age (Sclater *et al.* 1976b). Hence the difference between the estimates of heat flow through marginal basins and through the normal ocean basins was caused by the depression of the conductive heat flow through normal ocean basins by hydrothermal circulation, not by some heat generating process beneath the marginal basins.

This discussion is of particular importance in the Aegean. Jongsma (1974) has shown that the mean heat flux through the sea-floor is 2.1 $\mu\mathrm{cal\,cm^{-2}\,s^{-1}}$, or about twice that through the sea-floor south of the arc. This value is also substantially higher than the continental mean of 1.5 $\mu\mathrm{cal\,cm^{-2}\,s^{-1}}$. Makris (1975) has shown that the P-wave velocity beneath the Moho under the Aegean is 7.7 km/s, or substantially lower than that under stable regions. Since there is now no evidence that the high flow in regions like the Aegean is the result of processes beneath the lithosphere, both observations must be explained by some other mechanism. Similar low velocities are common beneath extensional zones with high heat flow. These observations are easily explained if the lithosphere of the Aegean has been stretched by a factor of two, as was argued in the previous section. To obtain the thermal structure and the surface temperature gradient it is necessary to know the temperature structure before stretching and the time which has elapsed since the thickness changed. The simplest geotherm has a constant gradient with depth: the one used here is shown in Fig. 20(a) and is similar to that proposed by Parsons & Sclater (1977) for old oceanic plate. Beneath a continent the temperature gradient must decrease with depth because of the contribution of radioactive heating, but if this effect is included the analysis below becomes more complicated without becoming more accurate. A number of different arguments suggest that the stretching started during the early Pliocene, perhaps 4 Myr ago. The calc-alkaline vulcanism associated with the present subduction zone is no more than 2.7 Myr old (Fytikas *et al.* 1976), and is distinct both in time and space from the Miocene and earlier vulcanism (Vilminot & Robert 1974; Fytikas *et al.* 1976). Mercier *et al.* (1976) have argued that the Aegean has been created by extension starting in the lower Pliocene about 5 Myr ago. These dates

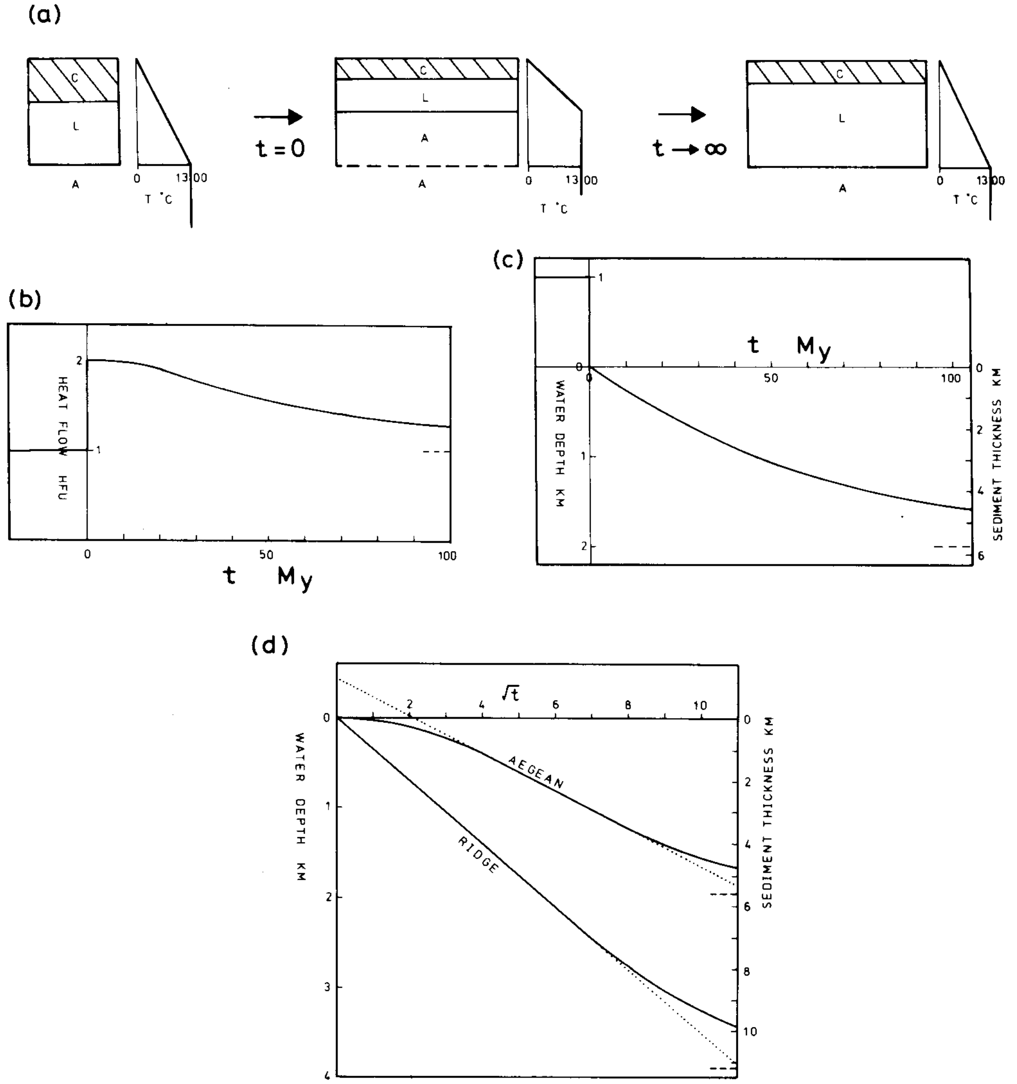

Figure 20. (a) Sketch to show the thermal structure of continental lithosphere, whose surface was originally 1 km above sea level, when extended by a factor of 2 at $t = 0$. The immediate result is to raise the boundary between the lithosphere and asthenosphere. This boundary gradually descends as heat is lost to the surface. The final thermal structure is the same as that before stretching. (b) The surface heat flow as a function of time, obtained from equation (4), HFU (1 HFU = 41.8 mW m^{-2}). The initial stretching doubles the heat flow, which decays, approximately exponentially, to its final value shown by the horizontal dashed line at the right. (c) Subsidence of the top of the lithosphere when there is no sedimentation (left side) and when the sedimentation rate is sufficiently fast to fill the subsiding region (right side), obtained from equation (6). Sediment density taken to be 2.5 gm cm^{-3}. The base level of the subsidence is shown by the horizontal dashed line on the right. (d) Subsidence of the stretched model, shown by the continuous line marked Aegean, plotted against $\sqrt{t}$, where t is in Myr. The dotted line shows a graphical extension of the straight-line portion of the curve. The ridge model of Parsons & Sclater (1977) is shown for comparison.

are consistent with the observations on land in both Greece and Turkey (Aubouin 1973; Brunn *et al.* 1976). Recently Lalechos & Savoyat (1977) have discovered Miocene evaporites in basins in the northern Aegean, and argue that the extension began somewhat earlier than other authors have suggested. However, all these estimated times are short compared with the thermal time constant of 60 Myr of a 120-km thick plate, and therefore do not affect the thermal structure.

The temperature of a rock element, therefore, cannot change appreciably during the extensional phase. Hence the present variation of temperature with depth is also a straight line, but the lithosphere is half as thick and the heat flow is twice that before extension. The lower half of the lithosphere has been replaced by mantle at the same temperature as the asthenosphere. This mechanism increases the temperature at shallow depths without forming sea-floor. Another explanation of these observations proposed by Makris (1976), requires the existence of a hot plume of rising mantle beneath the Aegean. Apart from the lack of independent evidence for its existence, it is hard to understand why such a plume should exist beneath the Aegean in particular, rather than elsewhere. The surface tectonics of the area are certainly different from that of surrounding regions, but it is difficult to understand why this should produce a hot plume from deep within the mantle.

If the crustal thickness of the Aegean has been reduced by a factor of two by stretching, about 200 km of motion must have occurred between Crete and Eurasia. If the extension began 3 Myr ago the relative velocity required is 7 cm/yr, or 3.5 cm/yr if the time is doubled. These velocity estimates are compatible with those in the previous section. Since any extension in the Aegean must be balanced by consumption along the Hellenic Arc, the length of the sinking slab must also be compatible with the model. The length of the slab dipping beneath Greece is about 200 km and has been produced by relative motion between the Aegean and African plates. Therefore this observation places an upper limit on the displacement between the Aegean and Eurasian plates of 200 km, and this limit will only be approached if the relative motion of the Aegean is fast compared with that between Africa and Eurasia.

An important consequence of the thermal model in Fig. 20 is that it gives rise to substantial subsidence as the lithosphere thickens and cools. The history of the subsidence is easily obtained from the cooling slab model (McKenzie 1967; Parsons & Sclater 1977) by expanding the temperature perturbation caused by the extension as a Fourier series. Since there are no lateral variations in temperature the conservation of energy requires

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} \tag{1}$$

where κ is the thermal diffusivity and z the vertical coordinate. If a is the thickness of the lithosphere before extension and T_0 the temperature at its base (1) can be simplified if dimensionless parameters are used, defined by

$$z = az', \qquad t = \frac{a^2}{\kappa} t', \qquad T = T_0 T'.$$

When (1) becomes

$$\frac{\partial T'}{\partial t'} = \frac{\partial^2 T'}{\partial z'^2}. \tag{2}$$

The solution to (2) with Fig. 20(a) as an initial condition is

$$T' = 1 - z' + \frac{4}{\pi^2} \sum_{l=0}^{\infty} \frac{(-1)^l}{(2l+1)^2} \exp[-(2l+1)^2 \pi^2 t'] \sin(2l+1)\pi z'. \quad (3)$$

Hence the surface temperature gradient is

$$\left(\frac{\partial T'}{\partial z'}\right)_{z'=1} = 1 + \frac{4}{\pi} \sum_{l=0}^{\infty} \frac{(-1)^l}{(2l+1)} \exp[-(2l+1)^2 \pi^2 t']. \quad (4)$$

The heat flow obtained from this expression is shown in Fig. 20(b). Terms other than $l = 0$ contribute little to the summation and are strongly affected by the details of the extension process. It is therefore sufficient to retain only the first term.

$$\left(\frac{\partial T'}{\partial z'}\right)_{z'=1} = 1 + \frac{4}{\pi} \exp(-\pi^2 t'). \quad (5)$$

This expression differs only by a factor of $2/\pi$ from the corresponding expression for the sea-floor spreading model (McKenzie 1967) even though the initial temperature conditions are very different. The resemblance is caused by the dominance of the first term in the summation, to which the near-surface structure contributes little.

The thermal contribution to the elevation Δe may be obtained by integrating (3) (Parsons & Sclater 1977)

$$\Delta e = \frac{a\rho_0 \alpha T_0}{(\rho_0 - \rho_w)} \times \frac{8}{\pi^3} \sum_{l=0}^{\infty} \frac{(-1)^l}{(2l+1)^3} \exp[-(2l+1)^2 \pi^2 t'] \quad (6)$$

where ρ_0 is the density of the mantle, ρ_w that of sea-water and α is the coefficient of thermal expansion. This expression is only approximately correct since it does not take account of the difference between the density and thermal expansion of crust and mantle rocks. The resulting subsidence is shown in Fig. 20(c). As before, the first term in the summation is dominant giving

$$\Delta e = \frac{a\rho_0 \alpha T_0}{(\rho_0 - \rho_w)} \frac{8}{\pi^3} \exp(-\pi^2 t'). \quad (7)$$

As for the expression for the heat flow, there is a difference of a factor of $2/\pi$ from the corresponding oceanic model. As Parsons & Sclater (1977) point out, (4) and (6) may also be written as functions of $\sqrt{t'}$. Substitution of their values for the parameters into (7) gives a value of 2 km for $\Delta e(0)$, and a time constant for the subsidence of about 60 Myr. If the sedimentation rate is sufficiently rapid to fill the trough as it develops then the total subsidence ΔE is amplified by compensation

$$\Delta E = \frac{\rho_0 - \rho_w}{\rho_0 - \rho_s} \Delta e \quad (8)$$

where ρ_s is the sediment density. Taking $\rho_s = 2.5$ gives $\Delta E = 5.75$ km, and a required initial sedimentation rate of about 8 cm/10^3 yr. Nesteroff, Wezel & Pautot (1973) found rates of 6 to 9 cm/10^3 yr for turbidite deposition in the Mediterranean Sea, and therefore the basin will fill as it subsides.

The subsidence curve is shown in Fig. 20(d) with Parsons & Sclater's (1977) ridge model for comparison. The principal differences between the two are that the linear portion of the

Aegean curve, between about 16 and 60 Myr, extrapolates to about +500 m, rather than passing through the origin, and both the total subsidence and rate of subsidence are less than that of the ridge model. However both curves depart from the linear relation between subsidence and age at approximately the same time.

The thermal model proposed above is similar to that of Sleep (1971), Sleep & Snell (1976) and Haxby, Turcotte & Bird (1976). The principal difference is that the present deformation of the Aegean suggests how the thermal disturbance which produces the subsidence arises. Unlike several of the proposals for generating such anomalies (Haxby *et al.* 1976), the asthenosphere plays a passive role, upwelling where stretching occurs. Hence the development of a basin can occur anywhere where a plate is stretched. This mechanism may be more common than frequently supposed, because the geometry of the normal faults discussed in the last section conceals much of the extension at the surface. Also the thermal anomaly prevents much of the subsidence which would otherwise occur due to isostatic compensation at the time of stretching. Only where good seismic reflection and refraction control is available will the true extent of the stretching be clear.

9 Driving forces

The discussion of the kinematics and thermal structure in the previous two sections ignored the dynamical problems involved in maintaining the observed motions. The two special features which stand out in Fig. 18 are the wide distribution of normal faulting and the close association of normal and thrust faults in northwestern Greece and Albania without any evidence of a sinking slab. There are other dynamical problems concerned with the force balances seaward of the Hellenic Arc and within the sinking slab, but these problems are harder to investigate in the Eastern Mediterranean than in the western Pacific, and will not be discussed here.

Since the momentum involved in all motions in the Aegean can be neglected, all such motions must be driven by forces now acting. There are three possible forces which could in principle maintain the observed motions: gravitational forces, forces on the boundaries of the region and forces on the base of the lithosphere. In a general way gravitational forces must be involved. They are the only source of the work required to maintain plate motions, and the similarity between the normal faults and landslips suggests that considerable gravitational energy is released during the extension. But it is difficult to believe that this release maintains the extension. All continents can release gravitational energy by becoming thinner, yet this behaviour is rare. Even in places where rapid and large variations in crustal thickness occur, such as between the Pacific Ocean and the Andes, normal faulting and extension is uncommon (Stauder 1975). In the Aegean region there is no great difference in thickness between the area south of the Hellenic Arc and the region undergoing extension. These arguments suggest that the extension is not the direct result of the local release of gravitational energy, but must be maintained by forces on the boundaries or base of the lithosphere which are generated by a release of potential energy elsewhere.

Boundary forces are probably important in continental tectonics. The westward motion of the Turkish plate is most easily maintained by boundary forces acting on the North and East Anatolian Faults (see I). Tapponnier & Molnar (1977) have also suggested that boundary forces can maintain similar motions on a larger scale further east. It is, however, difficult to understand how such forces can maintain the widespread extension observed in the Aegean area. Dewey & Sengor (1978) have attempted to do so, but it is hard to believe that blocks in northwestern Greece and Albania have sufficient strength to act in the manner they proposed. Many faults with variable trends are active, and the entire area is seismically active (Fig. 14).

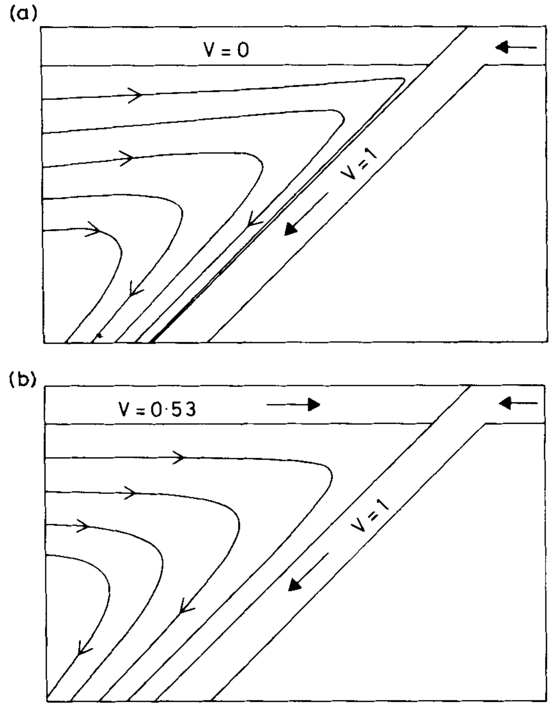

Figure 21. Flow in the wedge above the sinking slab when the plate behind the island arc is fixed with respect to the line of subduction, (a), and (b) when the upper surface above the sinking slab is allowed to move towards the line of subduction and is not acted on by any external force (see text).

The arguments above suggest that only forces acting on the base of the lithosphere can account for the observations. Because so little is known about the forces which maintain plate motions, the major constraints on such force systems must come from numerical and laboratory experiments on convecting systems. The observations require some mechanism by which the Aegean can be kept in tension and carried southward relative to Turkey. The change in mechanism from normal to thrusting on faults with similar strikes in Albania and northwestern Greece, and also in southwestern Turkey, requires rapid changes in the stress systems (Section 6), whereas the normal faulting normal to the arc does not (Section 5). In southwestern Turkey there is a seismically active sinking slab, but there is no evidence that such a slab exists beneath northwestern Greece and Albania. The similarity solution for the flow and stress behind a slab sinking through a viscous fluid of constant density is shown in Fig. 21. The solution in Fig. 21(a) is that obtained by McKenzie (1969a). The plate above the sinking slab is stationary with respect to the corner and the viscous forces on the base produce tension. If the stress is sufficient to break this plate, then the appropriate solution to use has zero tangential stress rather than zero velocity as a boundary condition at the surface. This solution is easily obtained by the same methods used by McKenzie (1969a):

$$\mathbf{v} = (v_r, v_\theta) = \left(\frac{d\Theta}{d\theta}, -\Theta\right) \tag{9}$$

where

$$\Theta = \frac{2V(\theta \sin\alpha \cos\theta - \alpha \cos\alpha \sin\theta)}{\sin 2\alpha - 2\alpha}. \tag{10}$$

v_r is the radial and v_θ the tangential velocity, V the radial velocity of the slab sinking at an angle α to the horizontal and θ is measured downward from the horizontal. The corresponding streamlines are shown in Fig. 21(b). A surprising feature of this solution is that the radial velocity on the upper surface is independent of r, the distance from the corner:

$$v_r(\theta = 0) = -\frac{2V(\sin\alpha - \alpha\cos\alpha)}{\sin 2\alpha - 2\alpha}. \qquad (11)$$

A plate moving with this velocity towards the corner will experience no stress on its base. A value of α of 45° gives $v_r = -0.53\,V$. The attractive feature of this solution is that it allows the extension associated with the subduction to occur at considerable distances behind the trench, even if the plate above the sinking slab is weak. In the Aegean region this is important because the extension is now taking place over a large area. The main difficulty with this model occurs if the motion of the material above the sinking slab is communicated directly to the plate above, as is shown in Fig. 21(b). The velocities shown are relative to the corner of the subducting plate, and would require extensive continental thickening beneath the southern margin of the Aegean. It is more likely that the flow in the asthenosphere is decoupled from the plate by a region of low viscosity.

The remaining problem concerns the deformation in northwestern Greece and Albania. The close relationship between the thrusting and normal faulting is similar to that in other parts of the Hellenic Arc. This connection suggests that there is a sinking zone in the mantle beneath the thrusting and upwelling further east. Such a flow could be maintained by a hot plume of material beneath the Aegean as Makris (1976) has suggested, or a cold sinking region further west. The problem with Makris' proposal is that upwelling plumes must originate in a hot boundary layer, either near a depth of 700 km or deeper. It is difficult to understand how surface motions could control the sites of such plumes, yet there appears to be a close relationship between the tectonics of northwestern Greece and Albania and that of the rest of the Hellenic Arc. A simpler model allows the lower part of the lithosphere to detach from the upper and to sink as a cold blob beneath the thrust zone. Similar motions are common in convecting fluids driven by heating from within (McKenzie, Roberts & Weiss 1974). Such an instability is easily produced by thrusting, which thickens the lithosphere. The lower part then detaches as a blob. The stress at the surface (Fig. 22) closely resembles that suggested by the fault plane solutions. The main difference is that, in the numerical experiments, the thrust zone is flanked on both sides by extension, whereas only one such zone is

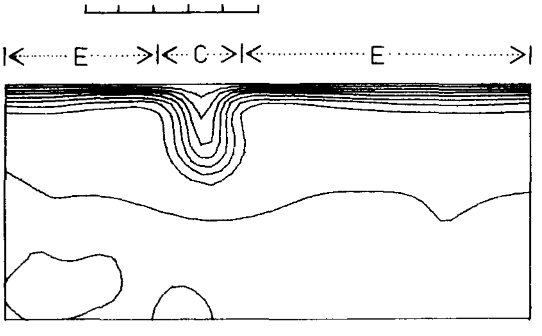

Figure 22. Isotherms obtained from a numerical model of convection in a fluid heated from within, showing the formation of a sinking blob of cold fluid by an instability of the upper boundary layer. Parts of the surface undergoing extension are marked E, those undergoing shortening are marked C. The subdivisions of the scale are at 100-km intervals.

observed in the Aegean. If, however, the strength of the Apulian block which exists beneath the northern part of the Ionian Sea is sufficient to prevent failure, then this difference can be explained. If there is a sinking blob beneath this area why is there no associated seismicity or vulcanism? The absence of earthquakes may be due to the temperature. McKenzie (1969a) showed that there was a close relationship between the temperature and the depth of the deepest shocks in sinking slabs. More recently Engdahl, Sleep & Lin (1977) have shown that the intermediate shocks beneath the Aleutians occur in the upper colder part of the plate (defined by the zone of anomalously high velocity). The intermediate shocks beneath the Aegean also appear to lie in the upper part of the plate (Gregersen 1977) though the control is not as good as in the Aleutians. The critical temperature is not well determined, but appears to be in the region of 500 to 700°C. Since the asthenosphere temperature is about 1300°C, sinking blobs can be both cold and aseismic. Furthermore, if the temperature is sufficiently high to cause elastic stresses to be relaxed by creep, the geometry of the cold sinking region will probably be different from that of a typical dipping slab.

The absence of vulcanism is less difficult to account for, since it is widely believed that calc-alkaline lavas are produced by partial melting of oceanic crust. There is no reason to believe that oceanic crust exists beneath northwestern Greece and Albania, and hence volcanoes should not be present. The cold blob is more likely to consist of mantle material cooled by the process discussed in the previous section than of true oceanic plate.

Arguments concerned with stresses and with the driving mechanism are always more speculative than those concerned with kinematics and thermal structure, and the ideas discussed in this section are no exception. Despite such warnings there is a suggestion that the surface motions in wide zones of continental deformation may directly reflect motions in the mantle below. The scale of the proposed motions beneath the Aegean is comparable to that of small-scale motions which are thought to exist beneath oceanic plates (Richter 1973; Parsons & McKenzie 1978). These ideas can be tested by travel-time observations such as those of Gregersen (1977), which should be able to detect high-velocity blobs in the mantle.

10 Discussion

The Aegean region is similar to other parts of the Alps and the Mediterranean, such as the Tyrrhenian Sea and the Pannonian Basin (Boccaletti *et al.* 1976) except that it is still seismically active. This difference is of great importance because fault plane solutions and the seismicity can be used to investigate the instantaneous deformation. By using the present deformation as a guide to that in the past, the problem of the mechanism which maintains the motions can be separated from the kinematic and thermal events. This is the same separation that was used with such success in plate tectonics. There is close geological resemblance between the Aegean Sea and the mountain belts of Greece and Turkey. The principal difference is that the Aegean continental crust is only about half as thick as that on either side. The stretching which has presumably produced this effect is still occurring in western Turkey and parts of Greece, though it has almost ceased in the southern part of the Aegean. The stretching is not confined to a small number of faults but occurs throughout large regions. The reason why the deformation should occur in this way is not understood. Because the dip of the normal faults involved decreases with depth, estimates of extension made from the dip and throw measured at the surface are always too small. Without seismic control or numerous drill holes reasonable estimates of the extension are not possible. One important consequence of the extension is to replace the lower part of the lithosphere with hot mantle, producing high heat flow and the slow thermal subsidence characteristic of sedimentary basins. The forces that maintain the observed motions are likely to result from flow associated with sinking slabs and from the cold lower parts of the lithosphere detaching and sinking as a blob through the mantle. Whether the descent of the cold material is marked by intermediate shocks depends on its temperature. These suggestions imply that the lower part of the lithosphere can be replaced, cooled and detached without generating oceanic crust and forming a spreading ridge. Lithosphere consisting of a thin (25 km) continental crust overlying mantle material appears to be more easily deformed than oceanic lithosphere and hence the surface motions reflect more closely the convective motions below than they do in true oceanic regions. Except where aseismic regions exist the motions are poorly described by plate tectonics. These ideas are strikingly similar to those proposed by the few tectonic geologists who accepted continental drift early this century (Argand 1924; Holmes 1929).

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References

Allen, C. R., 1975. Geological criteria for evaluating seismicity, *Bull. Geol. Soc. Am.*, **86**, 1041.

Alptekin, O., 1973. Focal mechanisms of earthquakes in western Turkey and their tectonic implication, *PhD thesis*, Faculty of New Mexico Institute for Mining and Technology.

Ambraseys, N. N., 1971. Value of historical records of earthquakes, *Nature*, **232**, 375.

Ambraseys, N. N., 1975. Studies in historical seismicity and tectonics, in *Geodynamics today, a review of the Earth's dynamic processes*, pp. 7–16, Royal Society, London.

Ambraseys, N. N. & Tchalenko, J. S., 1972. Seismotectonic aspects of the Gediz, Turkey, earthquake of March 1970, *Geophys. J. R. astr. Soc.*, **30**, 229.

Ambraseys, N. & Zatopek, Z., 1969. The Mudurnu Valley, West Anatolia, Turkey, earthquake of 22 July 1967, *Bull. seism. Soc. Am.*, **59**, 521.

Angelier, J., 1977. Sur l'evolution tectonique depuis le Miocéne Supérier d'un arc insulaire Méditerranéen: L'Arc Egéen, *Rev. geogr. phys. Geol. dyn.*, **19**, 271.

Argand, E., 1924. La tectonique de l'Asie, *C. R. Congr. Géol. Intern.*, 13e 1922 (Liege), pp. 169–371.

Arpat, E. & Bingol, E., 1969. The rift system of western Turkey: thoughts on its development, *Turkey Min. Res. Explor. Inst. Bull.*, **73**, 1,

Arsovsky, M., 1970. Contemporary tectonic properties of some active zones in Jugoslavia, *Bulg. Acad. Sci.*, **1**, 184.

Arsovsky, M. & Hadzievsky, D., 1970. Correlation between neotectonics and seismicity of Macedonia, *Tectonophys.*, **9**, 129.

Aubouin, J., 1973. Des tectoniques superposées et de leur signification par rapport aux modéles géophysiques: l'exemple des Dinarides; paleotectonique, tectonique, tarditectonique, néotectonique, *Bull. Soc. géol. Fr. (7)*, **15**, 426.

Aubouin, J., Bonneau, M., Davidson, J., Leboulenger, P., Metesco, S. & Zambetakis, A., 1976. Esquisse structurale de l'Arc égéen externes: des Dinarides aux Taurides, *Bull. Soc. géol. Fr. (7)*, **18**, 327.

Bernoulli, D., de Graciansky, P. C. & Monod, O., 1974. The extension of the Lycian Nappes (SW Turkey) into the southeastern Aegean Islands, *Eclo. Geol. Helv.*, **67**, 39–90.

Bingöl, E., 1976. Evolution géotectonique de l'Anatolie de l'ouest, *Bull. Soc. Geol. Fr. (7)*, **18**, 431.

Boccaletti, M., Horvath, F., Loddo, M., Mongelli, F. & Stegena, L., 1976. The Tyrrhenian and Pannonian Basins: a comparison of two Mediterranean interarc basins, *Tectonophys.*, **35**, 45.

Bott, M. H. P., 1976. Formation of sedimentary basins of graben type by extension of the continental crust, *Tectonophys.*, **36**, 77.

Brinkmann, R., 1976. *Geology of Turkey*, (F. E. Verlag, Stuttgart), Elsevier.

British Petroleum, 1971. *The geological results of petroleum exploration in western Greece*, Institute for Geology & Subsurface Research, Athens, No. 10.

Brunn, J. H., Argyriadis, I., Ricou, L. E., Poisson, A., Marcouz, J. & Graciansky, P. C., 1976. Elements majeurs de laison entre Taurides et Hellénides, *Bull. Soc. géol. Fr. (7)*, **18**, 481–497.

Crampin, S. & Ucer, S. B., 1975. The seismicity of the Marmara Sea region of Turkey, *Geophys. J. R. astr. Soc.*, **40**, 269.

Dewey, J. F. & Sengor, A. M. C., 1978. Aegean and surrounding regions: complex multi-plate and continuum tectonics in a convergent zone, *Bull. Geol. Soc. Am.*, in press.

Dewey, J. W., 1976. Seismicity of Northern Anatolia, *Bull. seism. Soc. Am.*, **66**, 843.

Dufaure, J.-J., 1965. Problèmes de Néotectonique dans le Péloponnèse, *Rév. Géogr. phys. Géol. dynam., Ser. 2*, **7**, 235.

Engdahl, E. R., Sleep, N. H. & Lin, M.-T., 1977. Plate effects in North Pacific subduction zones, *Tectonophys.*, **37**, 95.

Erinc, S., Bener, M., Sungur, K. & Göcmen, K., 1971. 12 Mayis 1971 Burdur Depremi, *IU Edebiyet Fakultesi Yayinlari, No. 1707.*

Fitch, T. J. & Muirhead, K. J., 1974. Depths to larger earthquakes associated with crustal loading, *Geophys. J. R. astr. Soc.*, **37**, 285.

Fytikas, M., Giuliani, O., Innocenti, F., Marinelli, G. & Mazzouli, R., 1976. Geochronological data on recent magmatism of the Aegean Sea, *Tectonophys.*, **31**, T29.

Gregersen, S., 1977. *P*-wave travel time residuals caused by a dipping plate in the Aegean Arc in Greece, *Tectonophys.*, **37**, 83.

Haxby, W. F., Turcotte, D. L. & Bird, J. M., 1976. Thermal and mechanical evolution of the Michigan Basin, *Tectonophys.*, **36**, 57.

Holmes, A., 1929. Radioactivity and Earth movements, *Geol. Soc. Glasgow Trans.*, **18**, 559.

I.F.P., 1966. *Etude Géologique de l'Epire (Grèce nord-occidentale)*, two volumes. Paris Editions Technip (L'institute de Geologie et researches du sous-sol-Athenes et l'Institut Français du Petrole – Mission Grèce).

Jongsma, D., 1974. Heat flow in the Aegean Sea, *Geophys. J. R. astr. Soc.*, **37**, 337.

Jongsma, D., 1975. A marine geophysical study of the Hellenic Arc, *PhD thesis*, Cambridge University.

Jongsma, D., Wissmann, G., Hinz, K. & Gardé, S., 1977. Seismic studies in the Cretan Sea. 2. The southern Aegean Sea: an extensional marginal basin without sea-floor spreading? 'Meteor' Forschungsergebnisse Reihe C, *No. 27, Geologie and Geophysik.*

Karig, D., 1971. Origin and development of the marginal basins in the western Pacific, *J. geophys. Res.*, **76**, 2542.

Ketin, I., 1966. 6 Ekim 1964 Manyas Deprimi Esnasinda Zeminde Meydana gelen Tansiyon Catlaklari, *Türk. Jeol. Kur. Bült.*, **10**, 44–51.

Ketin, I. & Roesli, F., 1953. Makroseismische Untersuchungen über das nordwestanatoligsche Beben vom 18 März 1953, *Eclo. Geol. Helv.*, **46**, 187–208.

Lalechos, N. & Savoyat, Ed., 1977. *La sedimentation neogene dans le fosse Nord Egeen* **VI**, Colloquium on the geology of the Aegean Region, Athens.

Longwell, C. R., 1945. Low angle normal faults in the Basin-and-Range Province, *Trans. Am. Geophys. Un.*, **26**, 107.

Makris, J., 1975. Crustal structure of the Aegean Sea and the Hellenides obtained from geophysical surveys, *J. Geophys.*, **41**, 441.

Makris, J., 1976. A dynamic model of the Hellenic Arc deduced from geophysical data, *Tectonophys.*, **36**, 339.

Makris, J., 1978. The crust and upper mantle of the Aegean region obtained from deep seismic soundings, *Tectonophys.*, in press.

Makris, J. & Vees, R., 1977. Crustal structure of the Central Aegean Sea and the Islands of Evvia and Crete, Greece, obtained by refraction seismic experiments, *J. Geophys.*, **42**, 329.

Maley, T. S. & Johnson, G. L., 1971. Morphology and structure of the Aegean Sea, *Deep-Sea Res.*, **18**, 109–122.

Mercier, J.-L., Carey, E., Phillip, H. & Sorel, D., 1976. La néotectonique plio-quaternaire de l'arc égéen externe et de la mer Égée et ses relations avec séismicité: *Bull. Soc. géol. Fr. (7)*, **18**, 159.

Molnar, P. & Tapponnier, P., 1975. Cenozoic tectonics of Asia: effects of a continental collision, *Science*, **189**, 419.

Morelli, C., Gantar, G. & Pisani, M., 1975. Bathymetry, gravity and magnetism in the Strait of Sicily and in the Ionian Sea, *Boll. Geofis. teor. appl.*, **17**, N65, 39.

Morelli, C., Pisani, M. & Gantar, C., 1975. Geophysical studies in the Aegean Sea and in the Eastern Mediterranean, *Boll. Geofis. teor. appl.*, **18**, 127.

Murawski, H., 1976. Raumproblem und Bewegungsablauf an listrischen Flächen, insbesondere bei Tiefenstörungen, *N. Jb. Geol. Paläont. Mh.*, **4**, 209.

McKenzie, D. P., 1967. Some remarks on heat flow and gravity anomalies, *J. geophys. Res.*, **72**, 6261.

McKenzie, D. P., 1969a. Speculations on the consequences and causes of plate motions, *Geophys. J. R. astr. Soc.*, **18**, 1.

McKenzie, D. P., 1969b. The relation between fault plane solutions for earthquakes and the directions of the principal stresses, *Bull. seism. Soc. Am.*, **59**, 591.

McKenzie, D. P., 1970. The plate tectonics of the Mediterranean region, *Nature*, **226**, 239–243.

McKenzie, D. P., 1972. Active tectonics of the Mediterranean Region, *Geophys. J. R. astr. Soc.*, **30**, 109.

McKenzie, D. P., Roberts, J. M. & Weiss, N. O., 1974. Convection in the Earth's mantle: towards a numerical simulation, *J. Fluid Mech.*, **62**, 465.

Needham, H. D., Le Pichon, X., Melguen, M., Pautot, G., Renard, V., Avedik, F. & Carré, D., 1973. North Aegean Sea Trough, 1972 Jean Charcot Cruise, *3e sess C.I.E.S.M. Athens*, **22**, 138.

Nesteroff, W. D., Wezel, F. C. & Pautot, G., 1973. *Summary of lithostratigraphic finds and problems*, Initial Reports of the Deep Sea Drilling Project, Vol. 13, US Government Printing Office, Washington, DC.

Ninkovich, D. & Hayes, J. D., 1972. Mediterranean Island Arcs and Origin of high potash volcanoes, *Earth planet. Sci. Lett.*, **16**, 331.

North, R. D., 1977. Seismic moment, source dimensions, and stresses associated with earthquakes in the Mediterranean and Middle East, *Geophys. J. R. astr. Soc.*, **48**, 137.

Nowroozi, A. A., 1972. Focal mechanisms of earthquakes of the Persian Plateau, Eastern Turkey, Caucasus and the Hindu–Kush regions, *Bull. seism. Soc. Am.*, **61**, 317.

Papazachos, B. C., 1973. Distribution of seismic foci in the Mediterranean and surrounding area and its tectonic implication, *Geophys. J. R. astr. Soc.*, **33**, 421.

Papazachos, B. C., 1976a. Seismotectonics of the Northern Aegean area, *Tectonophys.*, **33**, 199.

Papazachos, B. C., 1976b. Evidence of crustal shortening in the Northern Aegean region, *Boll. Geofis. teor. appl.*, **19**, 66.

Parsons, B. & Sclater, J. G., 1977. An analysis of the variation of ocean floor bathymetry and heat flow with age, *J. geophys. Res.*, **82**, 803.

Parsons, B. & McKenzie, D. P., 1978. Mantle convection and the thermal structure of the plates, *J. geophys. Res.*, in press.

Richter, C. F., 1958. *Elementary seismology*, W. H. Freeman, San Francisco.

Richter, F. M., 1973. Convection and the large scale circulation of the mantle, *J. geophys. Res.*, **78**, 8735.

Ritsema, A. R., 1974. *Earthquake mechanisms of the Balkan region*, UNDP/UNESCO report on Survey of the Seismicity of the Balkan Region, project REM/70/172.

Sclater, J. G., Crowe, J. & Anderson, R. N., 1976a. On the reliability of oceanic heat flow averages, *J. geophys. Res.*, **81**, 2997.

Sclater, J. G., Karig, D., Lawver, L. A. & Louden, K. 1976b. Heat flow, depth and crustal thickness of the marginal basins of the South Philippine Sea, *J. geophys. Res.*, **81**, 309.

Shirokova, E. I., 1972. Stress pattern and probable motion in the earthquake foci of the Asia–Mediterranean seismic belt, in *Elastic strain field of the Earth and mechanisms of earthquake sources*, eds Balakina, L. M. *et al.*, Seismology No. 8, Nauka Press, Moscow.

Sleep, N. H., 1971. Thermal effects of the formation of Atlantic continental margins by continental break up, *Geophys. J. R. astr. Soc.*, **24**, 325.

Sleep, N. H. & Snell, N. S., 1976. Thermal contraction and flexure of mid-continent and Atlantic marginal basins, *Geophys. J. R. astr. Soc.*, **45**, 125.

Stauder, W., 1975. Subduction of the Nazca plate under Peru as evidenced by focal mechanisms and by seismicity, *J. geophys. Res.*, **80**, 1053.

Stride, A. H., Belderson, R. H. & Kenyon, N. H., 1977. Evolving Miogeanticlines of the Eastern Mediterranean (Hellenic, Calabrian and Cyprus outer ridges), *Phil. Trans. R. Soc. Lond. A*, **284**, 255.

Sulstarova, E. & Kociaj, S., 1969. *Tërmeti i 30 nëndorit 1967 dhe brezi sizmogjen Vlorë-Dibër Buletini,* Univ. Shtetëror të Tiranës Ser. Shkencat Natyrore 2 65 (Tirana).

Tapponnier, P. & Molnar, P., 1977. Active faulting and tectonics in China, *J. geophys. Res.*, **82**, 2905.

Thompson, G. A., 1959. Gravity measurements between Hazen and Austin, Nevada: a study of Basin and Range structure, *J. geophys. Res.*, **64**, 217.

Vilminot, J.-C. & Robert, U., 1974. A propos des relations entre le volcanisme et la tectonique en mer Egee, *C.R. Acad. Sci. Ser. D*, **278**, 2099.

Woodside, J. M., 1975. Evolution of the eastern Mediterranean, *PhD thesis,* Cambridge University.

Wright, D., Jongsma, D., Stephen, M., Williams, S. & Matthews, D., 1975. Bathymetry of the eastern Mediterranean, *Boll. Geofis. teor. appl.*, **18**, 168.

Young, J. C., 1960. Structure and stratigraphy in North Central Scheel Creek range, in *Guidebook to the geology of the East Central Nevada,* eds Boettscher, J. W. & Sloan, W. W., Intermontane Ass. Pet. Geologists & E. Nevada Geol. Soc.